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Holocene environmental change in the Albertine Rift: Sediment-based evidence from the Virunga volcanoes

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Thesis submitted to the University of Dublin, Trinity College, for the Degree of Doctor in Philosophy

Discipline of Geography
School of Natural Sciences

The University of Dublin
Trinity College
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Declaration

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Thesis summary

The Albertine Rift is one of the most important conservation areas in tropical Africa, with exceptionally high levels of biodiversity (including rare and endemic montane flora and fauna) and high human population densities. Environmental changes in the Albertine Rift during the Holocene have been influenced both by long-term climate change and by human activity, although establishing clear cause-effect relationships is often problematic – particularly as this area is known to have had a long and complex history of human settlement. An understanding of the long-term drivers of environmental change in the area is important in order to provide a context for the possible effects of future climate change and other environmental perturbations.

This thesis uses sediment-based evidence from two crater sites in the Virunga volcanoes – a high-altitude area located along the borders between Uganda, Rwanda and the Democratic Republic of the Congo – to reconstruct the environmental history of the Albertine Rift during the mid- to late Holocene, and in particular to evaluate the main drivers of environmental change in the area. High-altitude sites such as these provide excellent potential to reconstruct past environmental changes over a wide surrounding area, in addition to reflecting within-site variations.

Sedimentary sequences were extracted from sites at the summits of two of the Virunga volcanoes: a crater swamp on Mt. Gahinga (3474 m) and a crater lake on Mt. Muhavura (4127 m). A combination of AMS\(^{14}\)C, \(^{210}\)Pb and \(^{137}\)Cs dating indicate that the sedimentary sequences extend back to ca. 8000 cal yrs BP and ca. 3000 cal yrs BP, respectively. Multi-proxy sediment-based analyses provide evidence for significant environmental changes in the area during the last several thousand years.

Pollen and charcoal records from the crater swamp on Mt. Gahinga show that climate changes linked to variations in monsoonal activity during the mid-Holocene have been an important driver of environmental change, particularly vegetation change, both at high-altitude sites and in the surrounding areas of the Albertine Rift. A fire-related decline in Ericaceous vegetation between ca. 5200 and 4900 cal yrs BP, followed by a later increase
in both high-altitude scrub and dry montane forest between ca. 4100 and 3700 cal yrs BP, provides evidence for vegetation responses to increased mid-Holocene aridity. However, the asynchronous timing of vegetation responses highlights the differential effects of climate on different vegetation types.

Human-induced environmental impact in the Albertine Rift is apparent in the sedimentary records from the sites in the Virunga volcanoes only within the second millennium AD, despite the long history of human occupation of the area. The sediment sequences from both the crater swamp on Mt. Gahinga and the crater lake on Mt. Muhavura indicate a significant forest clearance event at ca. 900 cal yrs BP, involving a reduction in lower montane forest taxa and increases in disturbance indicators. This period coincides with the onset of the Late Iron Age in the region, and thus highlights the increased role of human-induced environmental change since that time. Furthermore, significant changes in upper montane forest, possibly linked to changes in the fire regime associated with increased human activity, are also apparent from ca. 900 cal yrs BP. This highlights the widespread extent of human modification of the landscape within the last millennium, even in high-altitude areas previously thought pristine. The role of natural, long-term climate change as a major cause of environmental change in the Albertine Rift has thus been eclipsed within the last millennium by human-induced environmental effects.
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Chapter 1 Introduction

1.1 Environmental change in tropical high-altitude areas

The sensitivity of high-altitude areas to environmental changes is particularly apparent in the tropics, as is evident in the rapid retreat of tropical alpine glaciers due to recent climate changes (Jomelli et al., 2009; Thompson et al., 2009; Thompson et al., 2011). Further changes in the future can be expected. For example, climate models predict that high-altitude temperatures in the tropics will increase by up to 6°C by the end of the current century (twice as much as the 3°C increase predicted for average global sea-surface temperatures) (IPCC, 2007). In addition to climate change, other factors – particularly direct human pressures – also affect tropical high-altitude ecosystems. However, links between human societies and environmental change are complex, with feedbacks and interdependencies confounding clear cause-effect relationships. Moreover, the pairing of such powerfully resonant notions of 'climate change' and 'sensitive ecosystems' can potentially lead to over-simplifications – or narratives – in our understanding of complex environmental processes (Klein, 2002; Sheil and Wunder, 2002; Carswell, 2003). In order to provide a context for the possible effects of future climate changes and other environmental perturbations, an evidence-based approach is thus essential, including an understanding of how these environments have behaved in the past. This is particularly relevant in tropical high-altitude areas, where there is a relative dearth of long-term environmental monitoring compared with similar parts of more temperate latitudes (e.g. Battarbee et al., 2002; Koining et al., 2002; Haeberli et al., 2007).

The last several millennia in tropical Africa have been marked by a series of large-amplitude hydrological changes that often correlate on a broadly regional scale (Stager et al., 1997; Gasse, 2000), the effects of which have been detected to various extents in some high-altitude areas (e.g. Barker et al., 2001; Thompson et al., 2002; Street-Perrott et al., 2007). These widespread hydrological changes have generally been consistent with variations in monsoon strength, linked to orbitally-induced cycles of variation in insolation patterns (primarily due to the precession cycle) (Kutzbach, 1981; Verschuren et al., 2009). However, complex regional and local-scale interactions were also important, particularly
with regard to the often abrupt nature of the transitions between wet and dry episodes (Gasse, 2000; Castañeda et al., 2007).

Evidence of tropical African environmental history comes from a wide variety of sources, including documentary records, ice cores and lake sediments (Hamilton, 1982; Kiage and Liu, 2006). Instrumental weather records in tropical Africa usually extend back ca. 120 years, and some documentary records of weather-related information extend back to 1800 AD (Verschuren, 2004), although historical records of lake-level fluctuation for Lake Victoria, based on records from the upper Nile (the Rodah Nilometer), extend as far back as 700 AD (Nicholson, 1998). Ice cores can provide high-resolution proxy records of past climate change, and have been obtained in eastern Africa from Mt. Kilimanjaro (Thompson et al., 2002; Thompson et al., 2003). However, sedimentary records are the most widely distributed environmental archives in tropical Africa (Hamilton, 1982; Verschuren, 2003; 2004). Sediments accumulating in lakes and swamps provide a record of past ecosystem conditions and responses to environmental perturbations over long time-scales, and are particularly useful archives in tropical highland areas where the climate tends to be sufficiently wet to allow for the accumulation of dead organic matter (Hamilton and Taylor, 1986). Environmental reconstructions from sedimentary deposits are achieved through the use of sedimentary proxy data to infer changing ecosystem conditions. Vegetation history provides a particularly important dimension to the reconstruction of past environments in tropical Africa, as the extent and composition of vegetation types have undoubtedly responded to changing climates over long time-scales. However, vegetation changes in recent millennia have also been significantly influenced by human activity, leading to increased levels of complexity in the interpretation of palaeoecological records (Marchant and Hooghiemstra, 2004). The many feedbacks and inter-linkages between the various driving factors causing environmental change cause difficulties in ascribing an effect (as manifested in a palaeoenvironmental archive) to one particular cause. Multi-proxy sediment-based approaches provide scope for a more comprehensive interpretation of past environmental changes. Furthermore, many studies involving the reconstruction of various aspects of African history attempt to draw on more than one source of evidence in order to further inform and/or validate their interpretations. For
example, palaeoenvironmental evidence from sediment-based approaches can be placed in the context of documentary and oral evidence, historical linguistics, and archaeology (Campbell, 2004; Phillipson, 2005; Crumley, 2007) in order to gain a broader picture of social and environmental history.

The steep ecological gradients associated with montane regions cause high levels of biodiversity, particularly in tropical areas (Taylor, 1996). The area with which this thesis is concerned, the Albertine Rift in tropical eastern Africa, is associated with exceptionally high levels of biodiversity (Plumptre et al., 2007a), and is a focus of many conservation efforts at local, national and international scales, including transboundary conservation across three international borders (Uganda, Rwanda, and the Democratic Republic of the Congo) (Rainer et al., 2003; Owunji and Plumptre, 2010). Ecosystems associated with the Albertine Rift support rare montane flora and fauna, and are important socioeconomically in terms of ecosystem service provision and as sources of natural resources (Plumptre et al., 2007a). Records of environmental change from high-altitude sites are particularly useful as the immediate catchment is unlikely to have been directly modified by human activity. Such records can therefore reasonably be expected to archive past environmental changes over a wide surrounding area, in addition to reflecting variations much closer to the study site itself.

1.2 Thesis aims and research questions

This thesis aims to reconstruct the environmental history of a high-altitude area of the Albertine Rift during the mid- to late Holocene. In particular, the research aims to evaluate drivers of environmental change that have affected high-altitude areas during the Holocene, based largely on sedimentary evidence from two crater sites in the Virunga volcanoes. The following three research questions are addressed:

- To what extent have high-altitude sites in the Albertine Rift been affected by variations in monsoonal activity?

- Do high-altitude sediment records from the Albertine show evidence for increased human-induced environmental impact in recent millennia?
• What have been the effects of variations in climate, fire, volcanic activity and human activity on high-altitude sites in the Albertine Rift?

These research questions are addressed using multi-proxy, sedimentary evidence in the form of Holocene-aged sediment cores extracted from two high-altitude crater sites in the Virunga volcanoes, Albertine Rift. Multi-proxy studies provide a useful means of evaluating environmental changes across a range of spatial and temporal scales.

1.3 Thesis structure

This thesis begins with a background chapter (Chapter 2) describing past environmental and societal changes in eastern tropical Africa, in order to provide a context for the reconstruction of past environmental changes based on new multi-proxy sedimentary evidence from the Albertine Rift. Chapter 3 describes the Albertine Rift in more detail, in the context of tropical African highland areas, and provides a description of the Virunga volcanoes – the area within the Albertine Rift which is the particular focus of this study. The process of selecting suitable sites for the extraction of sediment cores, along with fieldwork methods to obtain the sediment cores and the subsequent laboratory methods applied in their analysis, are detailed in Chapter 4. Presentation of the results of sediment core extraction and analyses is divided into two chapters. Chapter 5 focuses on the establishment of chronological control for the sediment cores and the results of physical proxy measurements of the sediments. Chapter 6 presents the results of biological and chemical proxy analyses of the sediment cores. A summary and analysis of all results, and a comparison of the records from the two study sites, is presented in Chapter 7. Chapter 8 discusses of the results of this thesis in the context of the research questions, and places the findings in a broader context. Conclusions and concluding remarks are drawn in Chapter 9.
Chapter 2 Background to research

This chapter discusses the existing evidence for environmental change in tropical eastern Africa during the Holocene epoch, with particular reference to the latter part of the Holocene. This evidence presented is based primarily on records from terrestrial sediment cores, and is placed in the context of past societal changes, particularly for recent millennia. All dates are given in calibrated years (cal yrs BP, i.e. years before 1950 AD). Current climatic conditions and their main drivers in tropical Africa are briefly set out. Key points to emerge from this chapter are the spatial heterogeneity and complexity of palaeoenvironmental change in tropical eastern Africa on various timescales, and the many interacting processes driving environmental change – the impacts of which on high-altitude regions are still relatively poorly understood.

2.1 Tropical African climate: A brief context

Climate varies on all timescales, arising from several factors, both internal and external (Jones and Mann, 2004). In tropical Africa, temperature gradients are generally weak – diurnal variations of temperature are normally greater than seasonal differences. The main mechanism for the production of weather systems is airstream convergence triggering convection in the moist surface layer (Barry and Chorley, 2010). The trade wind systems of the two hemispheres converge along the Inter-Tropical Convergence Zone (ITCZ), a zone of low pressure characterised by convectional rainfall. However, convergence along the ITCZ is neither spatially nor temporally continuous, as the ITCZ migrates in response to seasonal changes in the location of maximum solar insolation (i.e. the ITCZ moves northwards from December to June, and southwards from June to December). Modern climate patterns in tropical Africa are largely governed by the seasonal migration of the ITCZ. The movement of the ITCZ causes, in turn, the seasonal migration of the equatorial rainfall belt, resulting in northern and southern zones of monsoonal climates with summer rains and winter drought, and double rainfall maxima near the equator (McGregor and Nieuwolt, 1998). The position and strength of the subtropical high-pressure cells (usually located between 20° and 30° north and south of the equator, and from which air flows towards the ITCZ), also has an important influence: these high-pressure cells move north
Fig. 2.1: Schematic of the general patterns of winds and pressure over Africa. Dotted lines indicate the ITCZ; dashed lines indicate the Congo Air Boundary. From Nicholson (1996).
during the northern hemisphere summer and south during the northern hemisphere winter. This contributes towards a northward shift in the location of the ITCZ over the continent of Africa during the northern hemisphere summer. Tracking of the thermal equator by the ITCZ is a major factor associated with the monsoon regimes of Asia, Africa and Australia (Barry and Chorley, 2010).

The climate of tropical Africa is not, however, influenced solely by the generally north-south movement of the ITCZ. Variations in climate across an east-west axis are linked to differences in surface and tropospheric temperatures, as well as to topographical features and to sea-surface temperatures (Gasse, 2000). Easterly and westerly air flows (the Indian Ocean monsoons and the Congo air stream, respectively) are separated by two main convergence zones – the ITCZ and the Congo Air Boundary (Nicholson, 1996). The Indian Ocean monsoons in eastern Africa are thermally stable and relatively dry, whereas the Congo air stream carries extremely moist and unstable air from the Atlantic (Nicholson, 1996). Rainfall fluctuations are also linked to El Niño-Southern Oscillation and to the Indian Ocean dipole (Gasse, 2000; Hastenrath, 2007; Marchant et al., 2007; Becker et al., 2010).

The distribution of highland areas alters regional temperature and moisture patterns: highlands often act as water towers for the surrounding lowlands (Taylor, 1996; Gasse, 2000). Furthermore, major variability in tropical climates can occur through local topographic effects on airflow. In particular, localised variations in precipitation related to aspect and altitude occur in highland areas (Taylor, 1996). In eastern and central Africa, the windward slopes of mountains between about 2000–3000 m are usually characterised by a marked maximum of cloud cover and rainfall, with a drier zone above (Street-Perrott and Perrott, 1993). Temperatures fall at an average rate of approximately 1.6°C per 300 m altitude increase, but this adiabatic lapse rate varies with topography, exposure and humidity (Hedberg, 1964; Taylor, 1996).

Orbital cycles (also known as Milankovitch cycles) are important long-term drivers of climate change on a global scale, and cause changes in the amount and seasonal distribution of incident solar radiation (insolation). The cycle involving the precession of
the equinoxes (the "wobble" of the earth's axis of rotation) is the shortest orbital cycle, with a periodicity of ca. 22000 years. Precession mainly affects the intensity of the seasons, and is particularly important at low latitudes, as it influences the amount of summer insolation received between 30°N and 30°S (McGregor and Nieuwolt, 1998; Clement et al., 2004). This in turn affects the strength of the monsoonal system: when summer insolation is lower, the differential heating of the land and oceans is not as great, leading to a restriction of monsoonal flows and less precipitation in the tropics (deMenocal et al., 2000b). During periods of higher summer insolation, the enhanced ocean-continent temperature and pressure contrast draws the monsoon winds farther inland and leads to intensified monsoon circulation (Verschuren et al., 2009).

The monsoonal circulation system is thus intricately linked to large-scale rainfall distribution in tropical Africa, particularly eastern Africa. Past variations in monsoon strength have had an important influence on hydrological changes in tropical Africa (Gasse, 2000). Milankovitch-induced insolation forcing undoubtedly plays an important role in driving long-term (millennial-scale) changes, but the African monsoon system is also affected (on sub-millennial scales) by tropical and North Atlantic Ocean circulation and sea-surface temperatures (Lamb et al., 1995; deMenocal et al., 2000b; Russell et al., 2003b; Schefuß et al., 2005). In central and eastern Africa, rainfall variability has often been linked to such forcing factors as the El Niño-Southern Oscillation (ENSO), the changing position of the ITCZ and the Indian Ocean dipole (e.g. Camberlin et al., 2001; Marchant et al., 2007; Russell et al., 2007; Tierney et al., 2008).

2.2 Holocene environmental change in tropical Africa

2.2.1 Early to mid-Holocene climate change: the African Humid Period

The base of the Holocene epoch has been formally defined as 11650 cal yrs BP (Walker et al., 2009). Prior to this, the late-glacial (ca. 15000 to 11650 cal yrs BP) was characterised by the widespread shrinkage and disappearance of the large continental ice-sheets that covered much of the northern hemisphere. In tropical Africa, the late-glacial constituted a series of abrupt transitions from the cold, arid conditions of the last glacial period towards more humid conditions, linked to changes in the global climate system (Gasse, 2000).
During the late-glacial and early Holocene, there was a significant increase in levels of humidity in tropical Africa, which is reflected in many lake-level records (e.g. Haberyan and Hecky, 1987; Talbot and Johannessen, 1992; Stager et al., 1997; Chalié and Gasse, 2003; Russell et al., 2003a; Marshall et al., 2009). Large areas that are now associated with aridity regularly received substantial rainfall: for example, the vast, currently arid expanses of the Sahara were fertile at this time, and a giant lake (Lake Mega-Chad) extended over a huge area in the Chad basin in the southern Sahara (Maley, 1977; Claussen and Gayler, 1997; Schuster et al., 2005). Oxygen isotope records from Mt. Kilimanjaro ice-cores indicate warmer and wetter conditions from ca. 11000 to 4000 cal yrs BP (Thompson et al., 2002). This is further corroborated by sediment-based evidence from alpine lakes on Mt. Kenya, where oxygen isotope ratios of diatom silica have yielded variations that are consistent with an increased lake moisture balance for much of the early Holocene (Barker et al., 2001). Furthermore, sediment records indicate significant vegetation changes during the late-glacial and at the onset of the Holocene: the cold, arid conditions of the last glacial were generally characterised by more xerophytic vegetation types than those that were widespread during the late-glacial and early Holocene (e.g. Lézine 1989; Jolly et al., 1997; 1998; Umer et al., 2007; Sinninghe Damsté et al., 2011). Time-transgressive forest re-establishment and re-expansion – particularly at higher altitudes – occurred in response to increased temperatures and precipitation towards the end of the late-glacial period (e.g. Taylor, 1990; 1992; Jolly et al., 1997; Taylor et al., 1999; Ryner et al., 2006; Umer et al., 2007). The timing of forest expansion varies from site to site, in some cases beginning before 14000 cal yrs BP (e.g. Taylor, 1990; Bonnefille et al., 1995), but almost always beginning by the time of Holocene onset (Ryner et al., 2006).

The early to mid-Holocene period of increased humidity in tropical Africa is known as the African Humid Period, during which time much of tropical Africa experienced a much warmer and wetter climate than at present (Gasse, 2000), although regional variability was also apparent (Bonnefille and Chalié, 2000). The main driver of this increased level of humidity was increasing levels of precipitation arising from the increased strength of monsoonal activity, largely due to orbitally-induced insolation forcing (Lézine 1989; Sirocko et al., 1996; Verschuren et al., 2009). Climate mechanisms were linked to sea-
surface temperatures in both the Atlantic and Indian oceans (Castañeda et al., 2007; Tierney et al., 2008), and an additional important factor was vegetation-atmosphere-ocean feedbacks, which further strengthened the monsoonal system through the amplification of land-sea contrast (Kutzbach and Street-Perrott, 1985; Ganopolski et al., 1998; Hély et al., 2009).

Evidence from throughout tropical Africa indicates a widespread shift to drier conditions during the mid-Holocene, involving a major decline in precipitation and a drop in lake and river levels, and thus marking the end of the African Humid Period. Many well-dated lake level records show evidence for this mid-Holocene aridity shift, the onset of which usually ranges from ca. 5800 to 5000 cal yrs BP, with minimum lake lowstands generally occurring between 4000 and 2000 cal yrs BP (Gasse and Street, 1978; Beuning et al., 1997; Lamb et al., 2000; Chalié and Gasse, 2002; Stager et al., 2003; Russell and Johnson, 2005). However, an apparent transition from humid to arid conditions is not widely synchronous, and in some cases does not occur until as late as 3200 cal yrs BP (Russell et al., 2003c). In the Sahara and Sahel, dust accumulating in lake sediments indicates that climate deterioration began shortly after 5000 cal yrs BP, with pronounced aridification evident by ca. 4100 cal yrs BP (Street-Perrott et al., 2000). The onset of aridity appears in many cases to have been rather rapid: for example, sediment records from Lake Edward, which is thought be representative of rainfall variability across much of tropical eastern Africa (Russell and Johnson, 2005), show an abrupt shift in the balance between precipitation and evaporation towards more evaporative conditions at 5200 cal yrs BP (Russell et al., 2003a). Oxygen isotope records in cores of ice from Mt. Kilimanjaro also indicate abrupt aridity shifts during the mid-Holocene, at ca. 5200 and 4000 cal yrs BP (Thompson et al., 2002).

Climate change during the mid-Holocene led to the onset of generally cooler and more arid conditions, with a widespread lowering of lake levels compared with the early Holocene (Street and Grove, 1976; Street-Perrott and Perrott, 1990; Sirocko et al., 1993; Marshall et al., 2009). This shift may be attributed in large part to a response to gradual precessionally driven changes in seasonal insolation, which led to a decline in the strength of the monsoonal system across tropical Africa during the mid-Holocene (Kutzbach and Street-Perrott, 1985). However, the asynchronous and often abrupt nature of the climate response
to gradual orbitally-induced insolation forcing suggests the operation of strongly non-linear feedback processes. Climate model studies invoke vegetation and ocean temperature feedbacks (particularly in the Sahara and subtropics) as likely mechanisms for this non-linear climate sensitivity (deMenocal et al., 2000a; Russell et al., 2003a).

The well-documented mid-Holocene shift in climate dynamics in tropical Africa should not, however, be envisaged as a spatially uniform step from a humid early Holocene to a drier and more variable late Holocene (Russell et al., 2003a). Several large, pronounced climatic oscillations have occurred in tropical Africa over the past 5000 years, the amplitude and timing of which are not always regionally synchronous. Drought events affected the Lake Edward basin at ca. 4000, 2000 and 900 cal yrs BP, which are likely to be linked to periods of reduced Indian Ocean monsoon intensity (Russell and Johnson, 2005). A significant lake-level lowstand is apparent in Lake Tanganyika at ca. 2000 cal yrs BP (Alin and Cohen, 2003), and in Lake Victoria at ca. 4100, 2500 and 1200 cal yrs BP (Stager et al., 2003; 2009). Arid episodes are evident in the Sahel from ca. 4300 cal yrs BP onwards (Street-Perrott et al., 2000; Kröpelin et al., 2008). Furthermore, the lack of a synchronous onset of aridity in the mid-Holocene indicates that the end of the African Humid Period did not consist of a single abrupt drying episode, but rather may have consisted of a complex stepwise trend towards a drier late Holocene, punctuated by many oscillations (Russell et al., 2003c; Marchant and Hooghiemstra, 2004).

Frequent short-term dry events during the late Holocene could be the reason why several lacustrine records are incomplete, with more arid conditions leading to desiccation and subsequent erosion of sediment sequences (deMenocal et al., 2000a; Marchant and Hooghiemstra, 2004). For those tropical African climate records that are continuous throughout the mid- to late Holocene, the several spatially variable sub-millennial climatic anomalies that are apparent (particularly in eastern Africa) suggest that orbital forcing is not the dominant pace-maker of decade- to century-scale climate variability during the mid- to late Holocene (Stager et al., 1997; Russell and Johnson, 2005). Furthermore, the periodicity evident in sedimentary records from continental parts of equatorial Africa is often similar to that recorded in depositional sequences from the Indian Ocean, and these similarities may indicate that low-frequency variability within the Indian Ocean (linked
also to ENSO events in the Pacific Ocean) dominates millennial-scale African climate shifts during the mid- to late Holocene (Russell et al., 2003b; Marchant et al., 2007).

Pollen data from many sites in tropical eastern Africa record significant vegetation changes during the mid-Holocene (e.g. Taylor, 1993; Taylor et al., 1999; Bonnefille and Chalié, 2000; Vincens et al., 2003; Umer et al., 2007), primarily in response to increasing aridity (Vincens et al., 2003). A shift towards vegetation types suited to drier conditions is widely evident in montane parts of eastern and central Africa between 5000 and 3000 cal yrs BP (e.g. Hamilton, 1982; Taylor, 1990; 1992; 1993; Umer et al., 2007). This particularly involves an expansion of Podocarpus, Juniperus, and Olea – the principal elements of relatively dry montane forest throughout much of the region (Lind and Morrison, 1974).

On Mt. Kenya, evidence for a vegetation response to aridification is shown by an expansion of C₄ grasses, a rise in Podocarpus, and reduced extent of other forest taxa between about 5000 and 4000 cal yrs BP (Olago, 2001; Rucina et al., 2009). Rucina et al. (2009) also found evidence for a dramatic increase in charcoal during this period on Mt. Kenya, which they surmise is most likely linked to a warmer, drier climate and increased fuel availability following increased forest growth. Furthermore, evidence from the Bale Mountains in Ethiopia shows an expansion of dry Afromontane forest (dwarf shrub and herbaceous vegetation) around 4500 cal yrs BP, and abrupt reduction in Ericaceous pollen from its early Holocene maximum (Umer et al., 2007).

Shifts towards drier vegetation types appear somewhat asynchronous from site to site. However, there is a widespread trend towards more xerophytic vegetation types after ca. 5000 cal yrs BP, which in some areas does not become apparent until around 3000 cal yrs BP. The stepwise nature of the climatic shifts towards increased aridity after the end of the African Humid Period, and the many drought periods that have affected parts of Africa during the late Holocene (as indicated by lake-level records), may explain the somewhat temporally and spatially variable vegetation response. The shift to drier conditions in tropical Africa may have been contemporaneous with a shift to wetter conditions in South America (Marchant and Hooghiemstra, 2004), which suggests an inter-related climatic cause which may partly be a result of stronger El Niño conditions (Lézine, 2007).
The overall picture of human settlement during the early to mid-Holocene in tropical Africa consists of widespread low-density Late Stone Age hunter-gatherer societies (using a wide range of increasingly specialised stone and bone tools), which may be seen as setting the stage for early settled food-producing societies (Vansina, 1995; Phillipson, 2005). Some semi-permanent settlements also existed around river and lake sites with rich fishing grounds (Brooks and Smith, 1987; Dale and Ashley, 2010; Prendergast, 2010). The increasing aridity of much of Africa from about 5000 cal yrs BP onwards (including the desiccation of the Sahara) is likely to have affected human settlement patterns, as many such river and lake sites were abandoned during the latter part of the Holocene, most likely as a response to increased aridity (Sutton, 1981).

2.2.2 Late Holocene: the increasing role of human impact

The separation and identification of the effects of various causes of environmental change becomes increasingly difficult during the latter part of the Holocene, as the influence of human activity becomes more widespread and intensive during this time. The development of agricultural production over the last few millennia, and the associated increase in population, has enabled humans to play a much more significant role in the shaping of the landscape. Palaeoecological evidence for vegetation change at several sites around 2500 cal yrs BP highlights a pattern of forest clearance (and possibly forest disturbance), often along with increased fire occurrence and/or intensity (Kendall, 1969; Taylor, 1992; Taylor and Marchant, 1995; Taylor et al., 1999; Vincens et al., 2003; Umer et al., 2007; Rucina et al., 2009). A climatic cause (or contribution) to these changes cannot be discounted, as this date coincides with the major lowstands (between 4000 and 2000 cal yrs BP) of several tropical African lakes. However, the timing also coincides with the transition from the Late Stone Age to the Early Iron Age in eastern Africa, during which some or all of the various communities occupying the region began to forge iron.

There is remarkable homogeneity among the archaeological evidence for early iron-working communities in eastern Africa, often associated with a characteristic pottery type (Urewe ceramics) and a sharp discontinuity in the archaeological record (Phillipson, 2005; Ashley, 2010). Most Early Iron Age Urewe sites date to around 2000 cal yrs BP, although some may be several centuries older (Schmidt and Childs, 1985; Clist, 1987): in a recent
review, Ashley (2010) gives the age range of Urewe ceramics as ca. 2500 to 1200 cal yrs BP. Urewe sites are generally located in areas of cultivation and abundant rainfall (Ashley, 2010), and the area around Lake Victoria is considered as one of the centres from which Early Iron Age technologies are likely to have spread to other areas of eastern and southern Africa (Phillipson, 2005). Early Iron Age settlement has long been associated with the expansion of Bantu-speaking farming populations (e.g. Oliver, 1966; Huffman, 1970; Phillipson, 1977), although the situation was most likely one of cross-cultural contact between different linguistic groups (Schoenbrun, 1993a; 1994; Feierman, 1995). Overall, the apparently widespread nature of primarily Bantu farming settlements, involving a diversified food-producing economy based on mixed-crop agriculture and stock-raising (Sutton, 1981; Schoenbrun, 1993a), indicates the potentially increased role of human impact on the landscape during the Early Iron Age.

The period from about 1000 cal yrs BP marks the onset of the Late Iron Age in many parts of eastern and central Africa, associated with widespread social, political and economic changes (Sutton, 1993; Schoenbrun, 1993b; Ashley, 2010). For example, in the Lake Victoria region, several large nucleated settlements were established, based on widespread cereal agriculture and cattle herding, with regional population increases and the emergence of a new ceramic culture (Sutton, 1998; Phillipson, 2005). Archaeological evidence for a number of particularly notable centralised settlements has emerged in what is now western Uganda. The earliest of these, at Ntusi, was occupied from at least the eleventh century AD, and appears to have been abandoned by the sixteenth century (Sutton, 1993). By the fifteenth century AD, larger polities began to emerge – at Bigo, Kibengo and Munsa. These larger polities were centred at settlements associated with earthworks that in some cases extended over several kilometres (Sutton, 1993; Robertshaw, 1994; Sutton, 1998). The earthworks were all abandoned towards the end of the seventeenth century AD, for reasons that are not fully understood (Taylor et al., 2000). Some oral traditions suggest this was a period of extensive political instability, migrations and famines; this instability could have been related to the drought conditions evidenced from lake-level records around this time (Robertshaw and Taylor, 2000; Taylor et al., 2000).
Late Iron Age settlement has been linked to further vegetation change occurring during the last millennium in eastern Africa, consisting of evidence for human-induced forest clearance and increased extent of cultivated land (e.g. Marchant and Taylor, 1998; Vincens et al., 2003; Ssemmanda et al., 2005; Lejju, 2009), and in some cases, changing limnological conditions (including eutrophication) (Russell et al., 2009) between ca. 1000 and 800 cal yrs BP. However, the complexity of factors influencing environmental change during the latter part of the Holocene, and particularly the most recent millennium, makes it difficult to attribute generalised causes on a regional scale. Climatic conditions also fluctuated during the past millennium: an arid phase between ca. 950 and 750 cal yrs BP is indicated in records from several sites – including Lake Naivasha, Kenya (Verschuren et al., 2000; Lamb et al., 2003), Lake Edward, Uganda (Russell and Johnson, 2007), Lake Masoko, Tanzania (Vincens et al., 2003), Lake Emakat, Tanzania (Ryner et al., 2008), and two crater lakes in western Uganda (Ryves et al., 2011). However, evidence for the Nile River discharge inferred from the Rodah Nilometer implies relatively wet conditions between about 950 and 650 cal yrs BP (Nicholson, 1998), and Ssemmanda et al. (2005) also attribute vegetation change in western Uganda during this time to wetter regional conditions. These different records suggest that there may have been strong regional gradients across the central African plateau at this time.

In eastern Africa, the periods of highest lake levels during the last millennium coincide with the three recent solar minima (Maunder, Spörer and Wolf) associated with the Little Ice Age (ca. 500–100 cal yrs BP), which suggests a close association between solar variability and eastern African water balance (Verschuren et al., 2000; Stager et al., 2005). The Little Ice Age, which was largely associated with the northern high latitudes, consisted of a complex set of regional climatic variations (Jones and Mann, 2004; Jones et al., 2009), and this complexity is also apparent in tropical latitudes. For example, within tropical Africa, there is evidence for a pronounced east-west rainfall gradient during the Little Ice Age, such that increased rainfall in the eastern arm of the East African Rift System (Verschuren et al., 2000; Stager et al., 2005) was synchronous with drought in the western arm (i.e. the Albertine Rift) (Russell and Johnson, 2005; Russell et al., 2007). This was most likely caused by a forcing mechanism linked to the shifting position of the ITCZ and
the interaction of El Niño-Southern Oscillation effects (Russell and Johnson, 2007), and highlights the possible linkages between high- and low-latitude climate variability. However, Ryves et al. (2011) suggest that wet conditions persisted in western Uganda during at least some of the Little Ice Age, indicating the complex spatial signature of regional climate changes during this time. The increased role of human impact during the past millennium is set against the backdrop of these regional climate variations: improved understanding of the relative roles of climate and human activity as drivers of environmental change in tropical Africa, including the Albertine Rift, is a major challenge, and meeting this challenge is a key aim of the current research.
Chapter 3 Study Area

This chapter provides a detailed context for the study area with which this thesis is concerned – the Albertine Rift in tropical eastern Africa, and more specifically, the Virunga volcanoes. The Great Rift Valley and its associated highland areas in eastern Africa are briefly discussed, followed by an account of the Albertine Rift and its surrounding area. A detailed description of the Virunga volcanoes is also provided.

3.1 Regional context: Tropical African Highland Areas

Eastern Africa contains the most significant clustering of highland areas in tropical Africa, most of which are associated with uplift or volcanic activity along the Great Rift Valley (see Fig. 3.1) – a continental rift system extending over more than 6000 km between northern Mozambique and the Red Sea (Schlüter, 1997). The Great Rift Valley is divided into two main arms: the Eastern (or Gregory) and Western (or Albertine) Rifts (Grove, 1983). Many areas of the African continent were unaffected by major global earth movements prior to the Palaeogene (ca. 65 to 23 Ma), resulting in a relatively subdued topography. However, since the latter part of the Palaeogene (ca. 33 Ma), and continuing through the Neogene (ca. 23 to 2.6 Ma), extensive uplift of the East African plateau has occurred, together with rifting, faulting and volcanism along the Great Rift Valley system (Hamilton, 1982; Sepulchre et al., 2006). Due to this tectonic uplift, much of eastern Africa lies at an altitude of >1200 m, but the highest elevations are those associated with volcanism and the graben shoulders of the two arms of the Rift Valley (Chorowicz, 2005).

Environment and climate of mountain systems are strongly influenced by latitude, altitude, topography, and aspect. In particular, tropical montane ecosystems are hotspots of biodiversity: temperature and available moisture can vary greatly over relatively small distances, producing steep altitudinal gradients in climatic conditions and in the composition and nature of ecological communities (Taylor, 1996). Tropical forest ecosystems at high altitudes are often considered more sensitive to climatic shifts than topographically subdued areas, and may therefore be particularly useful as barometers of environmental change (Diaz et al., 2003). A well-known feature of tropical Africa’s
mountains is the glaciers that currently occur in three areas: Mt. Kilimanjaro, Mt. Kenya and the Rwenzori Mountains. Traces of old glaciations are apparent in these three areas, as well as on Mt. Elgon and in the Aberdares (Hamilton, 1982). Rapid retreat of tropical alpine glaciers during the twentieth century – not only in Africa but also in South America and New Guinea – has been hailed as an indication that global warming poses a severe threat to tropical montane ecosystems (Gasse, 2002; Kaser et al., 2004). However, Kaser et al. (2010) highlight the complexity of processes contributing to ice fields on Mt. Kilimanjaro, and suggest that some of its ice fields have repeatedly disappeared and reappeared during the Holocene – a cyclicity that undermines the reliability of tropical glaciers as indicators of global warming or cooling.

Fig 3.1: Highland areas in eastern Africa. Redrawn and adapted from Lind and Morrison (1974).
3.1.1 Altitudinal vegetation belts

One of the most characteristic features of montane environments in tropical Africa is the occurrence of altitudinal vegetation belts and zones, in which similar vegetation communities are found in geographically distant upland areas (Lind and Morrison, 1974). As defined by Hedberg (1951), an altitudinal belt can be broadly traced on all mountains of sufficient height on a regional scale, whereas a zone is a mostly local altitudinal region. Hedberg (1951) distinguished three main vegetation belts on tropical (particularly eastern) African mountains: the Montane Forest belt, the Ericaceous belt, and the Alpine (or Afroalpine) belt. The following descriptions are based primarily on Hedberg’s (1951) classification. Fig. 3.2 illustrates the overall distribution of vegetation belts and zones for mountains in Uganda, Kenya and Tanzania.

The Montane Forest belt usually extends between 1700–2300 m and 3000–3300 m, and often contains three distinct zones: the montane rain-forest (or lower montane forest) zone, the bamboo zone and the Hagenia-Hypericum zone. All three zones are not always present, and they often intermingle. The lower montane forest zone is the lowest-altitude zone, dominated by broad-leaved hardwood trees and some conifers (particularly Podocarpus), and has often been heavily influenced by human activity. The bamboo zone consists primarily of large stands of mountain bamboo, Sinarundinaria alpina (K.Schum) C.S.Chao & Renvoize. The Hagenia-Hypericum zone forms the highest part of the belt, in which Hagenia abyssinica (Bruce) J.F. Gmel. and Hypericum revolutum Vahl are often the most important trees, with an undergrowth of tall herbs. The Ericaceous belt typically occurs from about 2600–3400 m up to an altitude of between 3500 and 4100 m, and can appear quite different on different mountains, but with considerable floristic agreement. It is usually dominated by arborescent or shrubby species of Philippia and Erica, and may also include some broad-leaved trees (such as Rapanea spp., Hypericum revolutum and Hagenia abyssinica), with Carex bogs occurring in level basins. The highest-altitude belt, the Afroalpine belt, is rather diverse on different mountains. The most typical plant communities are Dendrosenecio forest (including species of giant groundsel and giant lobelia), Helichrysum- and Alchemilla-scrub, and Carex tussocks.
However, any classification of vegetation belts and zones is in many ways arbitrary. All vegetation belts and zones may not be well-developed on every mountain, or may only occur on one aspect of the mountain (Osmaston, 2006). Water availability is a particularly important factor: the direction of the moisture-bearing air streams creates wetter and drier faces on the mountains, which can lead to distinct vegetation communities on different sides of the same mountain (Lind and Morrison, 1974). Steep hillsides and narrow valleys may also alter the local distribution of wetness and dryness. Other classifications of montane vegetation types tend to include a particular focus on dry and moist forest types, which is often useful when considering the floristic differences between mountains’ wetter and drier aspects (Hamilton, 1982). Intergradations of the different vegetation belts and zones are common, which makes the delimitation of the upper altitudinal limit of the vegetation belts very difficult and somewhat arbitrary. Furthermore, the original forest
vegetation has in many places been affected by human influence, resulting in replacement by cultivations and diverse types of secondary vegetation (Hedberg, 1951; Hamilton et al., 1986).

3.2 The Albertine Rift

The Albertine Rift is defined as extending from 30 km north of Lake Albert to the southern tip of Lake Tanganyika, including the valley, flanks of the escarpment and associated protected areas, and the range of species endemic to it (Fig. 3.3) (Owiunji et al., 2005; Plumptre et al., 2007a). It extends through parts of Uganda, Rwanda, the Democratic Republic of Congo (DRC), Burundi and Tanzania, and encompasses a wide altitudinal range, including numerous montane environments. Volcanism is an important feature of the Albertine Rift, with several regions of currently active volcanism, such as the southwestern part of the Virunga field in eastern DRC. However, not all of the montane environments in the Albertine Rift are volcanic in origin; for example, the Rwenzori Mountains are not volcanic, but are formed from a faulted, tilted and upthrust mass of Precambrian rock, which projects into the Albertine Rift (Osmaston, 1989).

The Albertine Rift encompasses a great diversity of habitats, and is a biodiversity hotspot of exceptionally high floral and faunal biodiversity. Its high species richness (consisting of both a high number of species and a large number of endemic species) mean that it is one of the most important regions for conservation in Africa. It contains more vertebrate species than any other region on the continent, and more endemic species of vertebrate than any other region on mainland Africa (Plumptre et al., 2007a; Owiunji and Plumptre, 2010). It has been designated an Endemic Bird Area by BirdLife International, an Ecoregion by the World Wildlife Fund, and a Biodiversity Hotspot by Conservation International. In particular, the high-altitude ecosystems of the Albertine Rift support rare endemic montane flora and fauna, and are also important socioeconomically as sources of natural resources (Plumptre et al., 2007a).
Fig. 3.3: The northern section of the Albertine Rift (between Lake Albert and Lake Tanganyika), and the interlacustrine region located on its eastern flanks (between Lakes Victoria and Kyoga on the east side and, to the west, the line of lakes associated with the Albertine Rift).
Plate 3.1: Steep topography and intensively cultivated land in southwest Uganda.

Plate 3.2: Tea plantation in southwest Uganda, with Bwindi-Impenetrable Forest in the background.
The Albertine Rift and the interlacustrine region (the area on the eastern flanks of the Albertine Rift, between a latitude of about 2°N and 5°S) are areas of enormous environmental diversity and rich social history. The Albertine Rift is particularly associated with steep, undulating topography and with rich, fertile volcanic soils (Chretien, 2003; Taylor et al., 2000). One of the interlacustrine region’s most distinctive physical traits is its altitude, which mostly ranges from about 1000 m to 2800 m, with summits around parts of the periphery reaching peaks of over 5000 m. The Albertine Rift and interlacustrine region have a long and complex history of human occupation, and currently support one of the highest rural population densities on the African continent, with up to 600-700 people per sq km in the central part of the region (southwest Uganda, Rwanda and Burundi and the adjacent areas of the DRC) (Plumptre et al., 2004). Land cover is characterised by a patchwork of intensively cultivated and grazed land, although moist broad-leaved and montane forest would have been formerly very extensive (White, 1983) (see Plates 3.1 and 3.2). Fragments of this forest still remain (e.g. Bwindi Impenetrable Forest), occurring mainly as relatively small, isolated outliers of the major rainforest block centred on the Congo basin (Taylor et al., 1999). The remaining areas of montane forest provide important watershed functions for adjacent areas (Plumptre et al., 2004). There are also areas of regenerating forest, often characterised by taxa such as Hagenia abyssinica, Macaranga capensis (Bail.) Benth. ex Sim, Neoboutonia macrocalyx Pax and Polyscias fulva (Hiern) Harms (Jolly et al., 1997). At higher altitudes, stands of mountain bamboo (Sinarundinaria alpina) are found, and above about 3000 m, Ericaceous scrub and montane grasslands dominate, although Hagenia trees are occasionally present. Burning and grazing activities have undoubtedly had a major effect on vegetation communities, and may even have affected the upper altitudinal limit of montane forests (Jolly et al., 1997). The Albertine Rift has been proposed as a core area of biological diversity for central Africa, which may mark the former limits of an important forest refuge during the cool, dry conditions associated with the peak of the last glacial period (Hamilton, 1984; Hamilton et al., 2001).
3.3 The Virunga volcanoes

The Virunga volcanoes (located along or close to the borders between Uganda, Rwanda and the DRC) comprise the southernmost currently active volcanic field within the Albertine Rift, and consist of numerous volcanic centres characterised by steep crater walls (Barifaijo et al., 2008). There are eight main peaks above an altitude of 3000 m, rising steeply from the surrounding landscape, and arranged into three groups (see Fig. 3.4). From east to west, these are: Mt. Muhavura (4,127 m), Mt. Gahinga (3,474 m) and Mt. Sabinyo (3,674 m), which stand close together on an east-west line along the Rwanda-Uganda border; Mt. Bisoke (3711 m), Mt. Karisimbi (4507 m) and Mt. Mikeno (4437 m), the latter of which is located in the DRC, while the two former sit astride the Rwanda-DRC border; and Mt. Nyiragongo (3470 m) and Mt. Nyamuragira (3058 m), which are located in the DRC (Plates 3.3–3.6).

Three contiguous protected areas straddle the borders of Uganda, Rwanda and the DRC – Mgahinga Gorilla National Park in Uganda, Volcanoes National Park in Rwanda, and the Mikeno Sector of Virunga National Park in the DRC. This constitutes the Virunga Conservation Area (sometimes known as the Virunga Massif), which is a relatively poorly demarcated area extending over ca. 430 km² in all three countries. Mgahinga Gorilla National Park was created in 1991 from land that had previously been gazetted as a Forest Reserve and Game Reserve, but the boundary fluctuated during the 1950s and 1960s, and was not always demarcated. Clearance of tracts of land (mostly lower montane forest) from the 1950s onward totally transformed the landscape by terracing and the introduction of crops and exotic tree species (Lejju, 2004). In 1991, however, the new boundary – which approximately followed the unmarked 1964 Game Reserve boundary – was demarcated, and all settlement and exploitation of resources in the park became by definition illegal (Adams and Infield, 2003). Virunga National Park (Mikeno sector) in the DRC and Volcanoes National Park in Rwanda were originally established in 1925 as part of Albert National Park, and have undergone several boundary changes since that time, including a period during which the parks were connected to Bwindi Impenetrable Forest by a buffer zone (Plumptre et al., 2007b).
Fig. 3.4: The Virunga volcanoes, and part of the surrounding area. 100 m contours above 2000 m are shown for the Virunga volcanoes.
Plate 3.3: View looking southwestwards towards the three volcanoes along the Uganda-Rwanda border. From left-right: Mt. Muhavura, Mt. Gahinga and (partially obscured by cloud) Mt. Sabinyo.


Plate 3.6: Westward view from the summit of Mt. Karisimbi, showing the two currently active volcanoes in the DRC: (l-r) Mt. Nyiragongo (with smoke rising from its lava lake-filled crater), and Mt. Nyamuragira (low-angle shield volcano). The sharp demarcation of the park boundary against intensively cultivated land can clearly be seen in the foreground. Photograph: David Taylor.
The Virunga Conservation Area is part of the broader Greater Virunga Landscape, which is an interconnected set of 11 protected areas along the three countries’ borders (Plumptre et al., 2007b). This landscape is a site of global importance for conservation, containing more endemic and threatened species, as well as more terrestrial vertebrate species, than any other site in Africa (Plumptre et al., 2007b). The Virunga Conservation Area alone contains 45–57% of the endemic vertebrate species found in the Albertine Rift (Owiunji et al., 2005). In the Virunga Conservation Area, the total number of plants recorded is 1265 (92 of which are endemic), and 294 bird species have been recorded (20 of which are endemic) (Owiunji et al., 2005). A total of 86 species of mammals have been recorded, of which 34 are large mammals, and 18 are endemic (Plumptre et al., 2003). Two of the large mammals, the mountain gorilla ({\em Gorilla beringei beringei}) and the Golden monkey ({\em Cercopithecus kandti}) are both endangered (www.iucnredlist.org). The volcanoes are home to more than half of the world’s remaining mountain gorillas (Kalpers and Lanjouw 1998), the remainder of which are found further north in the Albertine Rift, in Uganda’s Bwindi Impenetrable Forest. The most recent mountain gorilla census in the Virunga Conservation Area, carried out in the spring of 2010, recorded a total of 480 mountain gorillas (www.igcp.org), while a census in Bwindi Impenetrable Forest in 2006 recorded 320 mountain gorillas (McNeilage et al., 2006), bringing the total world population to 800 individuals. The International Gorilla Conservation Programme (IGBP) was established in 1991 to promote regional coordination and collaboration between the protected area authorities, with the aim of ensuring the conservation of the mountain gorilla and its habitat in all three countries (Rainer et al., 2003).

Of the eight main peaks of the Virunga volcanoes, only the two westernmost peaks – Mt. Nyiragongo and Mt. Nyamuragira – are currently active, while the six easternmost are considered dormant or possibly extinct. Mt. Nyiragongo and Mt. Nyamuragira have been active several times in the last few decades, and both volcanoes are surrounded by vast lava fields. Mt. Nyiragongo is a stratovolcano – a volcano built up of layers of lava alternating with beds of ash and other pyroclastics and with material eroded from higher slopes of the cone, and also known as a composite volcano (Allaby and Allaby, 1999). At least four of the other volcanoes – Mt. Muhavura, Mt. Gahinga, Mt. Karisimbi and Mt.
Bisoke – are also stratovolcanoes. Mt. Nyiragongo has steep slopes and a crater of about 2 km diameter, which usually contains a semi-permanent active lava lake (Plate 3.6). A major fissure eruption in 1977 drained the lava lake, resulting in a very fast-moving (up to 50 km/hr), destructive lava flow towards the town of Goma in the DRC (Tazieff, 1977; Tedesco, 2004). Another catastrophic lava flow issued from the volcano in 2002, reaching Goma and Lake Kivu, and destroying more than 120,000 people’s homes (Tedesco, 2007). These two major eruptions of Mt. Nyiragongo are thought to be of volcano-tectonic origin, related to tectonic activity in the Albertine Rift (Favalli et al., 2009). Mt. Nyamuragira is a low-angle shield volcano (Plate 3.6) which produces slow potassic-rich lava flows (Aoki et al., 1985). It has erupted at least 40 times since 1865, although most of the lava flows issue not from the summit (which contains a caldera 2 km in diameter) but from side-cones on the volcano’s flanks (Mavonga et al., 2006). Mt. Nyamuragira frequently emits huge quantities of ash during eruptions, leading to highly destructive ash falls (Tedesco, 2004).

Initial volcanism in this part of the Albertine Rift is likely to have begun around 12 Ma, associated with faulting and subsidence in the area (Ebinger, 1989). Several studies have obtained radiometric dates of volcanic rocks from the Virunga area and from throughout the Albertine Rift, and have found that the age of the earliest volcanic rocks decreases from north to south along the Albertine Rift axis: rocks from the Virunga area have been dated to between 11 and 9 Ma, while volcanic rocks from South Kivu, located farther to the south, yielded ages of up to 10 Ma, with a probable culmination of activity between about 6 and 5.5 Ma (Pasteels et al., 1989; Kampunzu et al., 1998). These dates are much later than the age of volcanism onset in the Eastern Rift at around 23 Ma, and suggest that the two arms of the Rift Valley are at different stages of development (Ebinger, 1989). Volcanic rocks are much less abundant in the Albertine Rift than in the Eastern Rift, and the rift grabens are narrower; this further confirms that the Western Rift is at a less mature stage than the Eastern Rift (Rogers et al., 1998). Volcanic activity in the Virunga area is most likely caused by a mantle plume underneath the Western Rift (Rogers et al., 1998), which contributes to the complexity of the magma series produced by the volcanoes (Vollmer and Norry, 1983).
While there is evidence to support the age of volcanism onset in the Virunga volcanoes, their recent eruptive history has proved much more difficult to determine. There is very little known on when the six dormant or extinct volcanoes were all last active, although it seems clear that they are of varying age. An eruption on the northern flank of Mt. Biseke, 11 km from the summit, occurred in 1957, lasting for less than two days and producing a 1 km long lava flow (Zana et al., 1989). Except for this short-lived eruption, only Mt. Nyiragongo and Mt. Nyamuragira have had eruptive episodes in the last 200 years (Mavonga et al., 2006). However, it is very unlikely that the Virunga volcanoes have been continuously active throughout the past 12 million years. Kampunzu et al. (1998) suggest that there have been two major volcanic episodes: (i) the initial Miocene episode, as indicated by the ages of between 12 and 8 Ma of the earliest volcanics, and (ii) the recent to present (Pliocene-Pleistocene) episode, from about 4 Ma onwards. These two episodes have produced distinctly different volcanic rocks (Kampunzu et al., 1998).

The main centre of volcanic activity in the Virunga volcanoes is likely to have generally moved from east to west over the past few million years (Aoki et al., 1985). However, the volcanoes’ overall morphology, as well as radiometric dating of their constituent volcanic rocks, can provide additional evidence for their relative – and actual – ages. Based on the volcanoes’ morphology, Mt. Sabinyo and Mt. Mikeno are generally believed to be the oldest, as they are the most deeply eroded and dissected (Rogers et al., 1998) (Plates 3.4 and 3.5). Mt. Karisimbi and Mt. Muhavura both have a rather youthful morphology (Plates 3.3 and 3.5), and are believed to have formed much more recently than Mt. Sabinyo and Mt. Mikeno (Rogers et al., 1998) – which, given that Mt. Muhavura is the easternmost volcano of the Virunga chain, highlights that the development of volcanic centres has been more complex than a simple east-west movement. Mt. Gahinga’s flat-topped summit and large, highly eroded and steep-sided crater suggest it was active much less recently than Mt. Muhavura (but more recently than Mt. Sabinyo). Similarly, Mt. Biseke is thought to be an older satellite to Mt. Karisimbi (Rogers et al., 1998).

Radiometric dating of volcanic rocks from the flanks of the volcanoes has suggested that most of the major volcanic structures of the Virunga volcanoes (with the exception of Mt. Mikeno) have formed within the last 200,000 years (Rogers et al., 1998). Potassium-Argon
(K-Ar) dates of rocks and minerals from Mt. Mikeno, the most eroded volcano, have yielded ages of up to 4.7 Ma (Guibert et al., 1975). Samples from Mt. Sabinyo, Mt. Karisimbi, and Mt. Muhavura have all given K-Ar and Argon-Argon (Ar-Ar) ages within the last 200,000 years (Kampunzu et al., 1998; Rogers et al., 1998), with only Sabinyo showing reliable ages of >100,000 years (Rogers et al., 1998).

A possible further strand of evidence comes from deposits of volcanic rocks and ash layers in late Quaternary sedimentary records from Muchoya swamp in southwest Uganda (Morrison, 1968; Taylor, 1988; Taylor et al., 1999). Morrison (1968) suggests that volcanic fragments pre-dating 13000 cal yrs BP could be from activity in the Virunga volcanoes (or possibly from the craters at Muko, in southwest Uganda). Similarly, Taylor (1988) presents evidence for three periods of volcanic depositions during the last ca. 42000 years – the most recent of which occurred between ca. 22000 and 14000 cal yrs BP. However, in terms of determining the potential source of these volcanic deposits, it proved difficult to chemically separate material from Mt. Gahinga, Mt. Muhavura, and the smaller ash cones around Kasheregyeni, Muko and Rubanda (Taylor, 1988). Given the K-Ar and Ar-Ar evidence that Mt. Muhavura (the closest volcano to Muchoya swamp) has erupted several times within the last 100,000 years (Rogers et al., 1998), it is certainly a distinct possibility that the volcanic deposits at Muchoya swamp represent an eruption of Mt. Muhavura – but conclusive evidence of this has proved elusive.

The impact of volcanic activity in shaping the landscape around the Virunga volcanoes has been marked. Lava flows have affected drainage and topography in the area, and have led to the development of rich, fertile volcanic soils in many places, highly suitable for cultivation (Snowden, 1933). Large volcanic rocks are often present in large numbers in the upper horizons of soils, which must be dug out before cultivation, and are then piled up in small piles of stones locally known as gahingas (from which Mt. Gahinga derives its name). Furthermore, the effect of past volcanism on vegetation in the area is likely to have been extensive, causing fires and ash falls, and periodic removal of existing vegetation cover (Taylor, 1988).
3.3.1 Vegetation zonation

The vegetation of the Virunga volcanoes shows a relatively marked altitudinal zonation, in common with other high-altitude sites in tropical Africa. However, there are many localised differences in vegetation distribution, depending on slope aspect, soil type, and human influence. The volcanoes often have sharp ridges radiating from their summits, separated by deep and narrow valleys and gullies, and the resulting contrast of soils and micro-climates causes a marked variation between the vegetation that occurs on the ridges and in the valleys. The high atmospheric humidity of the Virunga volcanoes is shown by the presence of many bryophytes (Hamilton, 1970). The zonation of vegetation on the three northeastern Virunga volcanoes (Mt. Sabinyo, Mt. Gahinga and Mt. Muhavura) is shown in Fig. 3.5, and a vertical representation of this (viewed from the north) is shown in Fig. 3.6. Photographs of some of the main vegetation types are shown in Plate 3.7.

![Vegetation zonation map](image_url)

**Fig. 3.5:** Vegetation zonation on Mt. Sabinyo, Mt. Gahinga and Mt. Muhavura. Adapted from Bygott and Hanly (1998) and 2003 Landsat data.
Plate 3.7: Vegetation types on the Virunga volcanoes: (a) lower montane forest; (b) bamboo; (c) Ericaceous belt; (d) Afroalpine belt.
Much of the vegetation at lower altitudes (including an extensive area within the current boundary of Mgahinga Gorilla National Park) has been influenced by human activity, primarily through clearance of the forest for agriculture. The montane forest below the bamboo (which usually extends up to an altitude of about 2500 m) has largely been cleared for cultivation, although some areas of secondary growth also exist. Areas of lower montane forest on the Virunga volcanoes mostly comprise relatively open woodland with widely spaced trees with dense ground flora, which is due both to human influence and to the high porosity of soils (Hamilton, 1970). Common trees in such areas include *Nuxia congesta* R.Br. ex Fresen, *Aguaria salicifolia* (Comm.ex Lam.) Hook.f., *Faurea saligna* Harvey, *Maesa lanceolata* Forssk., with an undergrowth of *Pycnostachys erici-rosenii* R.E. Fr. and species of *Carduus, Clutia, Cynoglossum, Kniphofia, Rumex* and *Vernonia* (Hedberg, 1951; Hamilton, 1970). On the Ugandan side of the volcanoes, several introduced species – including black wattle (*Acacia mearnsii* de Wild), *Eucalyptus* and *Cupressus* are reported to be increasing, although they still only comprise a small percentage of vegetation cover (Lejju, 1999; Owiumji et al., 2005).

A bamboo zone usually extends from ca. 2300 to 2600 m, but is not present on all of the volcanoes. Bamboo on the Virunga volcanoes tends to be thinner and smaller (only reaching heights of about 3 to 4 m) than on many other eastern African mountains, most
likely due to moisture deficiency caused by the shallow, porous volcanic soils (Hamilton, 1970). The bamboo zone is not well-developed on Mt. Muhavura, and does not occur at all on its eastern slopes, which are instead mostly covered by grassland. Where bamboo does occur, there is little undergrowth, but it is usually interspersed with other trees and shrubs, including *Hypericum, Dombeya, Pycnostachys erici-rosenii, Hagenia, and Vernonia* spp. (Hamilton, 1970).

In the upper part of the montane forest, above the bamboo (where it occurs), *Hagenia abyssinica* trees are widespread, along with smaller trees or shrubs of *Hypericum revolutum* and an undergrowth of tall herbs (mostly *Anthriscus silvestris* (L.) Hoffm.) (Hedberg, 1951). This forms a *Hagenia-Hypericum* zone that is extensive throughout the Virunga volcanoes (mainly between ca. 2600 and 3100 m), but which is often fragmented. *Rapanea rhododendroides* (Gilg) Mez is also a common tree in this zone. The *Hagenia-Hypericum* zone tends to not be well-developed on ridges that often radiate from the summits of the volcanoes, but occurs in the valleys and on hillslopes (Hamilton, 1970).

Ericaceous (heath) vegetation occurs between ca. 3000 and 3600 m, and is present on ridges. *Erica arborea* L. and *Philippia johnstonii* Schweinf. ex Engl. are common, especially on ridges (which tend to be drier), while the valleys at this altitude are often occupied by *Dendrosenecio erici-rosenii* (R.E.Fr. & T.C.E.Fr.) E.B.Knox and *Hypericum revolutum*. *Usnea* lichens are also common (Hedberg, 1951). Ericaceous vegetation is extensive on Mt. Muhavura, but is not well-developed on Mt. Gahinga.

Afroalpine vegetation generally occurs on the highest Virunga volcanoes from ca. 3600 m to the summit (Hedberg, 1951; Lind and Morrison, 1974), although it is also found above ca. 3300 m on Mt. Gahinga, and it occurs in gullies on Mt. Sabinyo and Mt. Muhavura down to ca. 3000 m. The main Afroalpine plant communities are dominated by *Dendrosenecio, Lobelia, Alchemilla* and *Helichrysum*.
Chapter 4 Research Methods

The methodological framework for this study is based on the extraction and analysis of sediment cores from high-altitude sites in the Albertine Rift, to provide evidence for palaeoenvironmental reconstruction. Following the identification of potentially suitable sampling sites, fieldwork to extract sediment cores was carried out, and a range of laboratory-based techniques were applied. The process of selecting sampling sites for core extraction, and the subsequent field and laboratory methods applied, are described and discussed in this chapter.

4.1 Sampling sites selection and sediment core extraction

In order to carry out sediment-based research, the first requirement upon which all other results depend is the retrieval of undisturbed sediment cores from a suitable sampling site – i.e. a site with (ideally) continuous deposition of sediment, from which it is possible to extract a sample that represents the vertical profile and reflects the entire basin as much as possible (Glew et al., 2001). For lake sediments, an undisturbed sample typically includes the sediment-water interface (i.e. a representative sample of the material that makes up the lake bed). This requires careful planning in terms of determining what type of sediment coring equipment is most appropriate for the site in question. No such requirement exists for swamp (i.e. peat) sediments. Transport of samples from the coring site is another important consideration, especially if laboratory storage facilities are a long distance from the coring site. Sediment samples should be kept dark and at low temperature to minimise microbial activity, which can influence determinations of organic compounds (Bengtsson and Enell, 1986).

The two easternmost Virunga volcanoes, Mt. Gahinga and Mt. Muhavura (located along the Uganda-Rwanda border), both have craters at their summits which contain sedimentary basins. The northern slopes of both volcanoes are located in Mgahinga Gorilla National Park in Uganda, and both sites can be accessed through this park, which makes it relatively easy to simultaneously arrange fieldwork at the two sites during the same period. A preliminary trip was undertaken to Mgahinga Gorilla National Park in March 2008, in
order to establish whether the crater sites at the summits of Mt. Gahinga and Mt. Muhavura were likely to be suitable for sediment coring fieldwork, and also to arrange fieldwork permission from the relevant Ugandan authorities. The crater swamp at the summit of Mt. Gahinga had been previously cored many years ago, but the cores had not been analysed (A. Hamilton, pers. comm.). This earlier successful extraction of cores, however, was a promising indication that there existed good potential for palaeoenvironmental work at this site. However, the crater lake at the summit of Mt. Muhavura did not appear to have ever been cored, or even had its depth measured. Determining the depth of the lake at the summit of Mt. Muhavura was thus important, in order to effectively plan subsequent coring missions. Preliminary depth measurements at the Mt. Muhavura crater lake indicated that the maximum depth of the lake was ca. 1.6 m. This made the site an extremely suitable candidate for coring from the side of an inflatable boat using a Livingstone piston corer, which can be used only in relatively shallow water conditions.

4.1.1 Mt. Gahinga crater swamp

The flat-topped summit of Mt. Gahinga (3474 m) comprises a large crater occupied by an elliptoid-shaped swamp, with steep-sided walls covered in dense Afroalpine vegetation (Plate 4.1). The width of the swamp is about 90 m, and it extends lengthways for approximately 200 m. Vegetation on the bog surface is dominated by several species of Sphagnum moss, along with large patches of Eriocaulon volkensii Engl., with Alchemilla johnstonii Oliv. and Romulea congoensis Bég. on the bog margins (Hamilton, 1970).

Fieldwork to extract sediment cores from the crater swamp on Mt. Gahinga was carried out in July 2008. The most suitable location for coring was determined by carrying out an initial depth-testing transect across the swamp, in order to establish the area of deepest sedimentation (Plate 4.2). This area – as long as it is at some distance from any channels or apparent surface disturbance – is likely to yield the longest, most continuous sediment record at the site. Sediment cores from the swamp were extracted in vertical one-metre sections using a modified square-rod Livingstone piston corer. Contiguous one-metre cores were extracted from the same coring hole. A Russian (D-section) corer was used to obtain the topmost section of sediment. All sediment cores were extruded into half-cylindrical
Plate 4.1: Crater swamp at the summit of Mt. Gahinga. X and Y mark the approximate end points of the depth-testing transect. Distance between X and Y is ca. 90 m.

Plate 4.2: Depth-testing transect prior to coring at the Mt. Gahinga crater swamp. Photograph: Stephen Taylor.
pieces of plastic guttering that had been lined with plastic film, and were wrapped and transported to Trinity College Dublin in whole-core (i.e. one-metre) sections, where they were placed in a cold store room at 2°C to await laboratory processing. The cores were later described using the Troels-Smith system (Troels-Smith, 1955; Aaby and Berglund, 1986), sub-sampled into 1-cm sections, and placed into labelled zip-lock bags. The outermost layer of sediment was removed before sub-sampling to minimise contamination.

4.1.2 Mt. Muhavura crater lake

Mt. Muhavura (4127 m) is shaped as an almost perfect cone, with a number of ridges radiating from its summit, separated by narrow valleys. The crater lake at the summit of Mt. Muhavura is a small, approximately circular, shallow lake (diameter ca. 25 m; maximum depth 1.6 m) surrounded by grasses and sparse Afroalpine vegetation in the shallow-sided crater (Plate 4.3).

Fieldwork to extract sediment cores from the crater lake on Mt. Muhavura was carried out in July and August 2008. Ropes were secured in two perpendicular directions across the lake, and a two-way depth-testing transect of the lake (using a hand-held echosounder and depth-testing rods) was carried out in order to determine the location of (i) greatest water depth and (ii) greatest sediment depth. This was used as a basis for deciding on the most suitable sites within the lake for core extraction. Sediment cores from the lake were extracted in vertical one-metre sections from two coring locations (MUH1 and MUH2) using a modified square-rod Livingstone piston corer, operated from an anchored inflatable boat secured to a transect rope (Plate 4.4). Cores of overlapping depth were obtained at each coring site. Casing to mark the coring hole was not used (due to the small size of the lake and its shallow depth of water). Sediment cores extracted from the same coring site were assumed to represent overlapping sections of the same sediment sequence, and can be correlated by depth accordingly. However, between-core section correlation of sediments was also based on obvious matching features in the sediment cores themselves. Sediment age was also taken into account, as determined by radiocarbon dating (see Section 4.2.3), when cross-correlating sediments from different sections of the core sequence.
Plate 4.3: Crater lake at the summit of Mt. Muhavura. Diameter is ca. 25 m. Photograph: David Taylor.

All sediment cores were extruded into half-cylindrical pieces of plastic guttering lined with plastic film (Plate 4.4). The cores obtained from the first coring site (MUH1) were described soon after extraction using the Troels-Smith system (Troels-Smith, 1955; Aaby and Berglund, 1986) and subsampled into 1-cm sections. The outermost layer of sediment was removed before sub-sampling in order to minimise contamination. Samples were placed into labelled zip-lock bags, and then double-bagged before transport to Trinity College Dublin. The cores obtained from the second coring site (MUH2) were wrapped and transported to Trinity College Dublin in whole-core (i.e. one-metre) sections, and were later described and sub-sampled into 1-cm sections following the above method. All cores and samples were placed in a cold store room at 2°C.

A Renberg gravity corer was used to obtain a short core of the topmost sediment, which was visually inspected to ensure that the sediment-water interface had been sampled with minimal disturbance. This core was sub-sampled in the field immediately following collection at 0.5 cm intervals for the first 5 cm, and at 1 cm intervals thereafter. Samples were placed into labelled zip-lock bags for transport back to Trinity College Dublin, where they were placed in a cold store room at 2°C.

4.1.3 Corrections for core shortening

Shortening of cores is a frequent occurrence with open-barrel type equipment (such as a Livingstone corer), whereby the length of sediment recovered is less than the penetrated depth of the coring barrel (Blomqvist, 1985; Glew et al., 2001). This is also sometimes known as core compression, although it does not consist of true compression, but rather of the progressive thinning of material (i.e. of sedimentary layers) due to frictional and deformation forces pushing sediment outwards as the core is pushed deeper (Glew et al., 2001). Fig. 4.1 illustrates this effect.

In this study, as a modified square-rod Livingstone piston corer was used to obtain all sediment cores except the uppermost cores from the two sample sites, core shortening must be accounted for. Each core of sediment extracted was assumed to represent the same length as the depth of penetration of the coring rods during each respective coring drive. If the sediment core length was less than the depth of coring rod penetration (which was, in
most cases, one metre), then a correction was applied based on the assumption that the rate of shortening of the core was the same throughout the length of the coring barrel (cf. Fig. 4.1).

**Undisturbed sediment**

**Core**

![Fig. 4.1: Effect of core shortening (core compression) for a hypothetical sediment core, in which layers in the sediment are progressively thinned down the core. Redrawn from Glew et al. (2001).](image)

### 4.2 Chronological control of sediment cores

The establishment of reliable sediment chronologies is a crucial aspect of any palaeoenvironmental investigation. A variety of radiometric techniques, based on the measurement of radionuclides, is available for producing an age-depth profile for a sedimentary sequence, which allows the timing and rate of past environmental changes to be determined. A radionuclide is an atom (known as the parent isotope) with an unstable nucleus that undergoes exponential radioactive decay to a daughter isotope by emitting or receiving sub-atomic particles. The time taken for half the original number of atoms to decay is termed one half-life, and radiometric dating involves comparing the number of parent and daughter isotopes in order to establish how many half-lives have passed since
decay began (i.e. in the case of sediment cores, the time since sediment deposition). In this study, core chronologies have been established by a combination of (i) $^{210}$Pb and $^{137}$Cs dating of the uppermost sediments and (ii) radiocarbon (AMS $^{14}$C) dating of plant macrofossils from throughout the sequences.

4.2.1 $^{210}$Pb and $^{137}$Cs radiometric dating

The analysis of short-lived isotopes of lead ($^{210}$Pb; half-life 22.3 years) provides an important method of dating recent sediments (i.e. for the last ca. 100–150 years) (Appleby, 2001). $^{210}$Pb is part of the $^{238}$U (Uranium-238) decay chain, which occurs naturally in many sediments, rocks and soils (Fig. 4.2). The parent isotope of $^{210}$Pb in the series is $^{226}$Ra (Radium-226); this decays initially to $^{222}$Rn (Radon-222) (a gaseous isotope), which then decays through a series of short-lived intermediate radionuclides to $^{210}$Pb. However, a fraction of the gaseous $^{222}$Rn atoms escape to the atmosphere by diffusion from soils and weathering, leading to the formation of atmospheric $^{210}$Pb. Atmospheric $^{210}$Pb has a relatively short residence time, becoming readily attached to airborne particulate material and removed by washout and dry deposition. The total amount of $^{210}$Pb in sediments is therefore a combination of $^{210}$Pb that has been derived from atmospheric fallout (unsupported $^{210}$Pb) and $^{210}$Pb that is in equilibrium with the in situ $^{226}$Ra (supported $^{210}$Pb) (Appleby, 2001).

\[ \begin{array}{cccccc}
238^\text{U} & \rightarrow & 226^\text{Ra} & \rightarrow & 222^\text{Rn} & \rightarrow & 210^\text{Pb} & \rightarrow & 206^\text{Pb} \\
4.51 \times 10^\text{y} & 1602 \text{y} & 3.82 \text{d} & 22.26 \text{y} & 138.4 \text{d} \\
\end{array} \]

**Fig. 4.2:** Uranium decay series, showing the principal radionuclides concerned with the production of $^{210}$Pb, and their radioactive half-lives (y = years; d = days). Dotted arrows indicate that intermediate radionuclides are not shown. Adapted from Appleby (2001).

Deciding on the appropriate dating model is important, as the dating model is determined by sedimentary conditions at the coring site. There are two simple models for $^{210}$Pb analysis, commonly referred to as the CRS (constant rate of $^{210}$Pb supply) model and CIC (constant initial concentration) model (Appleby and Oldfield, 1978). There are many situations in which neither of the simple models is valid, due to mixing of the surface
sediment by physical or biological processes, or to variations in the $^{210}\text{Pb}$ supply due to changes in the pattern of sediment focusing. To improve the $^{210}\text{Pb}$ model for a sediment sequence, independent chronological markers can be provided by the measurement of artificial radionuclides from nuclear fallout materials, such as $^{137}\text{Cs}$ (Caesium-137). Fallout of $^{137}\text{Cs}$ on a global scale began in 1954, and peaked in 1963 shortly after the test-ban treaty (Appleby, 2001). Lesser peaks also occur regionally, the most frequently referred to in the northern hemisphere being Chernobyl, dated to 1986.

In this study, sediment samples from the uppermost sediment core from the crater lake on Mt. Muhavura (extracted using a Renberg corer) were analysed for $^{210}\text{Pb}$, $^{226}\text{Ra}$ and $^{137}\text{Cs}$ by direct gamma assay in the Bloomsbury Environment Institute at University College London, using an ORTEC HPGe GWL series well-type coaxial low background intrinsic germanium detector. $^{210}\text{Pb}$ was determined via its gamma emissions at 46.5 keV, and $^{226}\text{Ra}$ by the 295 keV and 352 keV gamma rays emitted by its daughter isotope $^{214}\text{Pb}$ following storage for three weeks in sealed containers to allow radioactive equilibration. $^{137}\text{Cs}$ was measured by its emissions at 662 keV. The absolute efficiencies of the detector were determined using calibrated sources and sediment samples of known activity. Corrections were made for the effect of self-absorption of low energy gamma rays within the sample.

### 4.2.2 Radiocarbon dating

Radiocarbon dating is a widely used radiometric dating technique for obtaining the age of carbon-containing materials, and is a highly useful method for dating lake and swamp sediments from the late Quaternary (Björck and Wohlfarth, 2001). The radiocarbon technique is based on the decay of the radioactive isotope of carbon ($^{14}\text{C}$, or radiocarbon). $^{14}\text{C}$ is a cosmogenic nuclide which continually forms in the upper atmosphere by the interaction of neutrons (produced by cosmic rays) with nitrogen ($^{14}\text{N}$) atoms. $^{14}\text{C}$ atoms combine with oxygen to form $^{14}\text{CO}_2$, which mixes throughout the atmosphere, dissolves in the oceans, and enters the biosphere, where it is absorbed by all living organisms (Bowman, 1990). On death an organism stops absorbing carbon, and the $^{14}\text{C}$ radioactively decays to $^{14}\text{N}$ at a known rate. Hence, determining the age (i.e. date of death) of a sample involves measuring the amount of residual radiocarbon activity in a dead organism or in material derived from that organism. The half-life value of $5568 \pm 30$ years, as determined
by Libby (1955), is used by convention (ages are quoted as 'conventional radiocarbon ages'), although a more accurate measurement of 5730 ± 40 years was obtained by Godwin (1962). Radiocarbon dating has been successfully applied back as far as 50,000 years ago (Reimer et al., 2009). Accelerator mass spectrometry (AMS) is a method of sensitively measuring radiocarbon activity in very small samples, and has enabled precise dating of specific components of sediment, such as plant macrofossils (Jull, 2007).

The timescale for conventional radiocarbon ages (i.e. $^{14}$C years) is not the same as calendar ages. This is due both to natural variations in the amount of radiocarbon production in the upper atmosphere, and also to the incorrect (Libby) half-life. Conventional radiocarbon ages should therefore be converted to calendar ages by a process known as calibration (usually reported in calibrated years before 1950 AD). Calibration curves are created based on measurements of radiocarbon in samples that have been independently dated (e.g., tree-rings, corals, and annually laminated sediments) (Reimer et al., 2009). As the radiocarbon timescale (i.e. the calibration curve) is non-linear, calibration of even very precise radiocarbon measurements can result in complex, multi-modal age distributions, which vary depending on the shape of the calibration curve during that period. Evaluating the probability structure of multi-modal age distributions can be problematic. The uncertainty range associated with a conventional radiocarbon date is an indication of the precision (not accuracy) of the laboratory radiocarbon measurement, and is Gaussian in distribution, unlike the complex multi-modal probability distributions of calibrated radiocarbon dates.

In this study, a total of 21 AMS radiocarbon dates were obtained for cored sediments from the two study sites core sequences. Plant macrofossils (mosses) extracted from the cores were rinsed in deionised water, dried at 50°C, and packaged for transport. Four initial rangefinder samples were analysed by Beta Analytic Inc., Florida, and the remaining 17 samples were analysed at the 14CHRONO Centre, Queen's University Belfast. All conventional radiocarbon dates were analysed with 2σ errors. Calibration of radiocarbon ages was carried out against IntCal09 (Reimer et al., 2009) using the program OxCal v. 4.1.7 (http://c14.arch.ox.ac.uk/oxcal.html) (Bronk Ramsey, 2009).
4.2.3 Production of age models

Age models were produced for the core sequences from both sites, based primarily on the $^{14}$C dates, but also incorporating the results of the $^{210}$Pb dating model for the crater lake site. The age models aimed to find mathematically a representative set of possible ages for each depth point in the sedimentary sequences. The next section will describe the steps taken in the production of age models for both sites, and will also discuss the procedure followed to correlate the sequences of cores from the crater lake on Mt. Muhavura.

One approach to dealing with the multi-modal probability structure of radiocarbon dates involves using Bayesian probability, which provides a more sophisticated statistical tool for analysing such complex data. Bayesian probability allows for the inclusion of prior knowledge in the analysis of a set of data measurements – in the case of a sediment sequence, this means that stratigraphical context and succession (i.e. that age should increase with depth) can be included as a factor in the analysis of a set of calibrated radiocarbon dates (Blockley et al., 2004). Using Bayes’ theorem, the calibrated radiocarbon date for a sample (the likelihood) can be combined with the depth of a sample in a sediment sequence (the prior knowledge) to generate the posterior probability distribution which incorporates both the measured and prior information (Bronk Ramsey, 2008). The complex mathematics involved in Bayesian calculation cautions against a 'black box' approach (Pilcher, 2005: 68) that produces results with high precision, but low accuracy (Steier and Rom, 2000).

In the current research, the computer program OxCal v. 4.1.7 was used to apply a series of Poisson-process (or P-sequence) deposition models to the data. Such models allow for non-uniform deposition rates throughout the sediment sequence (i.e. they allow for fluctuations in the deposition rate) by viewing the sediment deposition as discrete events or increments, with a user-defined parameter $k$ as the number of increments per unit length. The larger the value of $k$, the smaller the increments, and the closer the model approximates to a constant deposition model. The level of $k$ chosen for a model will determine the rigidity of the model to fluctuations in the deposition rate. An agreement index $A$ gives the relative likelihood of the posterior from the model to that of a null model. As $k$ increases, less variability in deposition rate is allowed for, and there will usually be a
value at which all values of $A$ are above 60% (the usual threshold for acceptable $A$) (Bronk Ramsey, 2008). The $P$-sequence models for the core sequences from both study sites in the current research were run with initial values of $k = 0.1$, which was gradually increased in increments of 0.1 until the overall $A$ fell below 60%.

A further issue related to the overlapping sediment cores from the crater lake on Mt. Muhavura. During sediment coring, as outlined in Section 4.1.2, overlapping sediment cores were assumed to represent replicate sections of the same sequence, and could be correlated by depth accordingly. However, overlapping cores may not always correlate exactly according to depth alone: as casing was not used to mark the coring hole, successive sediment cores are unlikely to have been extracted from exactly the same location, but rather from slightly offset locations. This could lead to slight differences in the relative sediment depths obtained. Correlation of cores was therefore not solely based on the measured sediment depth during extraction, but also on (i) the stratigraphic and physical proxy data and (ii) the sediment age (as determined by radiocarbon dating). The stratigraphic and physical proxy data were used as a basis for determining two possible options by which the cores could be correlated, and Bayesian age models provided a way to evaluate the extent to which the radiocarbon dates supported each option. Based on these methods, the most suitable way to correlate the sediment cores was identified.

### 4.3 Physical proxies

#### 4.3.1 LOI analysis

Measurement of the percentage loss on ignition (%LOI) is a widely used method to estimate the organic content of sediments (Dean, 1974; Boyle, 2004), and also allows for the calculation of percentage dry weight and wet density of the sediment. Information on organic matter content can provide useful information relating to sediment inputs to the lake or swamp basin (Bengtsson and Enell, 1986). %LOI is also often a useful tool for the stratigraphical alignment of sediment cores from the same basin (Dearing, 1986).

An adaptation of the procedure outlined by Dean (1974) was followed, whereby 1 cm$^3$ of wet sediment was dried at 60°C for 48 hours and then ignited in a furnace at 550°C for
four hours. %LOI, % dry weight and wet density were measured at at least 8 cm intervals for all cores from the crater swamp on Mt. Gahinga, and at at least 2 cm intervals for all cores from the crater lake on Mt. Muhavura. Repeat measurements were carried out for all core samples analysed from the Mt. Muhavura sequence. In cases where the %LOI of the two measurements had a difference of more than 5%, a third replicate sample was analysed.

4.3.2 Magnetic susceptibility

Magnetic susceptibility is a measurement of the ease with which a material can be magnetised, and can be used as an approximation of the concentration of magnetic minerals in sediments (Thompson and Oldfield, 1986). The magnetic properties of sediments can be used as a basis for their characterisation, and can also be used (in conjunction with other sedimentary information) to aid correlation between different sediment cores from the same basin, and to identify common stratigraphical features in cores from different sites (Nowaczyk, 2001). Magnetic susceptibility varies with the content of iron-bearing minerals, which are common in nearly all rocks and their corresponding weathering products (Zolitschka et al., 2001). Magnetic susceptibility can therefore be regarded as a proxy for minerogenic contributions to the sediment, and can provide an indication of allochthonous input to the sedimentary basin, via terrestrial inwash or aeolian deposition. Furthermore, sediment magnetic properties are likely to be indicative of specific sources and processes, and variations in magnetic susceptibility (one such magnetic property of sediments) tend to be linked to other proxy indicators of palaeoecological and minerogenic changes (Thompson and Oldfield, 1986).

Whole-core scanning for magnetic susceptibility is a non-destructive method used to collect high-resolution data from intact sediment cores. Furthermore, whole-core scanning provides a way to gather information that can greatly aid core correlation, as stratigraphical features that may not be easily visible in organic-rich sediment cores can be highlighted. In this study, magnetic susceptibility measurements were obtained for all whole-core sections using a Geotek Multi-Sensor Core Logger (MSCL) at the National University of Ireland, Maynooth. As the sensor only penetrates approximately 1 cm into the sediment surface, magnetic susceptibility was measured on two opposite surfaces of the sediment cores in
order to obtain averaged values for each depth. Measurements were made at a resolution of 0.5 cm, with the exception of the six lowermost metres of sediment from the crater swamp on Mt. Gahinga, which were measured at a resolution of 1 cm. Appropriate adjustments were made for core shortening, where applicable.

Whole-core measurements were not possible for the set of sediment cores from the crater lake on Mt. Muhavura that were subsampled in the field prior to transportation back to Ireland (MUH1). Magnetic susceptibility measurements were instead obtained using a hand-held Bartington magnetic susceptibility meter at the National University of Ireland, Maynooth. Several samples from sediment cores for which whole-core measurements had previously been taken were also measured using the hand-held meter, in order to confirm the comparability of results from the two methods.

4.4 Biological and chemical proxies

4.4.1 Pollen analysis

Palynology, the study of fossil pollen and spores, provides a means of reconstructing past vegetation communities and of determining vegetation response to past environmental changes, whether climate- or human-induced (Bennett and Willis, 2001). Unless otherwise specified, the term pollen will be used henceforth to refer to both pollen and spores. Pollen preserved in sediments can be sourced both from the area directly surrounding the sedimentary environment and from distal locations (Moore et al., 1991), and can therefore be used to assess both local and regional vegetation changes, on a range of time-scales. Palynological evidence is particularly useful when combined with other proxy sources of palaeoenvironmental information, as this may allow the causes and drivers of vegetation change to be more fully assessed.

High-altitude sites provide the potential to reconstruct vegetation for a wide surrounding area, as pollen usually tends to be transported from low to high altitudes (Flenley, 1973; Solomon and Silkworth, 1986). Long-distance transport of pollen may be particularly facilitated where vegetation close to the sediment sampling site is relatively sparse or open (Hamilton, 1972; Marchant and Taylor, 2000). Hamilton (1972), in an analysis of pollen
grains in surface samples from a range of altitudes in highland Uganda, found that by far the majority of the long-distance pollen came from lower and not higher altitude vegetation. This may be due to lower pollen production by high-altitude plants and greater production by some forest taxa (Hamilton, 1972). The long-distance transport of forest tree pollen is also facilitated by the higher release into the atmosphere of tree pollen relative to that for herbaceous taxa closer to the ground. Meteorological conditions may also be important: on eastern African mountains, winds tend to blow up-slope during the day and down-slope at night (Flenley, 1973). This could influence pollen transport and deposition in two ways: (i) daytime is when most anemophilous plants tend to release pollen, resulting in its transport up-slope during the day; and (ii) frequent evening precipitation at high altitudes would help to wash the daytime pollen into sedimentary environments at high altitudes, or it could be deposited in the still evening air (Flenley, 1973).

Determining the vegetation reflected by a given pollen assemblage thus depends on an understanding of the relationship between (i) the production, dispersal and transport of pollen from the parent vegetation source and (ii) the deposition of pollen in a sedimentary environment. In montane areas of eastern Africa, including several sites in western Uganda, studies have been carried out to examine the relationship between pollen deposited in surface sediment samples and the surrounding vegetation (e.g. Hamilton, 1970; 1972; Bonnefille and Vincens, 1977; Marchant, 1997; Vincens et al., 1997; Marchant and Taylor, 2000; Umer et al., 2007), and such information can be used as an interpretative aid in analysing and discussing fossil pollen diagrams. Hamilton (1982) classified pollen inputs to sedimentary basins according to three pathways: locally sourced pollen is indicative of aqueous and sedimentary features close to the sample sites; non-locally sourced pollen is representative of the climate, land use and microclimate within the catchment for a sample site; and distal pollen reflects a wide range of regional conditions. For high-altitude sites, a distinction can be made between high-altitude (i.e. Ericaceous and Afroalpine belt) pollen and long-distance pollen. Following the terminology of Hamilton (1972), the long-distance element consists of all pollen grains that have actually been transported upwards from below the Ericaceous and Afroalpine belts, whereas the long-distance component consists of all pollen grains which can be identified.
as definitely transported upwards from below the Ericaceous and Afroalpine belts (i.e. if there are no corresponding Ericaceous or Afroalpine species). Pollen types can thus be classified based on their dispersal ability (Hamilton, 1972) – also referred to as ‘relative export ability’ (Flenley, 1973: 137). Low relative export ability refers to pollen types only present in samples collected within vegetation in which it occurs; moderate relative export ability refers to pollen types which are also present in samples collected in adjacent vegetation; and high relative export ability refers to pollen which occurs in all or almost all samples, wherever collected (Hamilton, 1972; Flenley, 1973; 1979). If a pollen type with low relative export ability is found abundantly in sediments, then it can be inferred that the parent taxon occurred in the vicinity of the sediment sampling site.

In this study, samples from cores from both study sites were prepared for pollen analysis following the standard procedure outlined in Faegri and Iverson (1989). To enable the determination of absolute pollen values, a known number of Lycopodium spores were added to each sample. Samples were mounted onto glass slides in silicone oil and sealed beneath coverslips, and pollen identification and enumeration was carried out at x 400 magnification on a Leica DMLS light microscope (in Trinity College Dublin) and a Leitz microscope (in the National Museums of Kenya, Nairobi). At least 500 pollen grains and spores were identified and counted for each sample, in accordance with the recommendations of Hamilton (1972) and Marchant (1997). Each slide was counted up to the edge of the cover-slip, so as to counteract possible differential pollen migration within the slide, and the ‘edge effect’ (Brookes and Thomas, 1967). The identification of pollen grains was based on comparisons with reference material (at the Department of Geography, Trinity College Dublin and the Palynology Department, National Museums of Kenya, Nairobi) and with several literature-based sources relating to eastern Africa and other African regions (including Hamilton, 1970; Bonnefille, 1971a; 1971b; Bonnefille and Riollet, 1980; Hamilton, 1982; Taylor, 1988; Marchant, 1997), and the African Pollen Database (http://medias.obs-mip.fr/apd).

The first set of pollen percentages was calculated based on the total pollen sum, excluding unidentified and damaged grains. All values are therefore relative to that of all the other taxa in the pollen sum. Pollen taxa were grouped according to whether they are associated
mainly with montane forest tree taxa, scrub/herbaceous taxa, other arboreal taxa, local/aquatic taxa, or unclassified taxa. An assessment was made of the dominance of local/aquatic taxa, and this group was excluded where appropriate.

A second set of pollen percentage values was calculated based on the long-distance and high-altitude pollen totals. The classification of long-distance and high-altitude pollen followed the findings of Hamilton (1972) for Mt. Elgon and the Rwenzori Mountains, and was further supplemented by vegetation information specific to the Virunga volcanoes (based on Hedberg, 1951; Hamilton, 1972; Owuunji et al., 2005; M. Popp, pers. comm.). Percentage values were calculated as a percentage of the respective long-distance or high-altitude sum, thus highlighting each taxon's abundance relative to other long-distance or high-altitude taxa. Some of the pollen types included in the high-altitude pollen group can also be produced by taxa from outside the Ericaceous and Afroalpine belts (e.g. Poaceae pollen), and some of the high-altitude pollen may therefore have been transported to the crater sites from outside those two belts. However, only pollen types that could be identified as definitely not coming from Ericaceous and Afroalpine vegetation were placed in the long-distance group (following Hamilton, 1972).

Pollen data were also expressed as concentrations (in $10^3$ grains per gram dry weight of sediment), in order to evaluate absolute pollen abundance, in addition to the relative abundances shown by the percentage data. Total concentration of pollen was calculated, in addition to concentration values for each pollen type. Total pollen accumulation rate (in $10^3$ grains per cm$^2$ per year) was also calculated. Concentration and accumulation rate calculation methods are summarised in Table 4.1. Sediment accumulation rates were calculated by linear interpolation between the mid-points of consecutive pairs of radiocarbon date age ranges determined by the age models.

In order to facilitate description, comparison and discussion of the pollen data, each core sequence was divided into pollen zones. Numerical methods provide a useful way of dividing a pollen sequence into zones based solely on specified mathematical criteria (Birks and Gordon, 1985). Pollen zones were defined numerically based on total pollen percentage data, using the program TILIA v. 1.5.12. A constrained incremental sum of
Table 4.1: Calculations used to obtain pollen concentrations and accumulation rates. Lyco added = number of Lycopodium spores added; Lyco counted = number of Lycopodium spores counted for each pollen count; Pollen counted = number of pollen grains counted (applied both for each taxa, and also for the totals, and for each category).

<table>
<thead>
<tr>
<th>Measurement</th>
<th>Unit</th>
<th>Formula</th>
</tr>
</thead>
<tbody>
<tr>
<td>% dry weight (%DW)</td>
<td>%</td>
<td></td>
</tr>
<tr>
<td>Wet bulk density</td>
<td>g cm⁻³</td>
<td></td>
</tr>
<tr>
<td>Dry bulk density</td>
<td>g cm⁻³</td>
<td>(%DW x wet bulk density) / 100</td>
</tr>
<tr>
<td>Accumulation rate</td>
<td>cm yr⁻¹</td>
<td>Sample depth / Sample age in yrs BP</td>
</tr>
<tr>
<td>Dry mass accumulation rate</td>
<td>g cm⁻³ yr⁻¹</td>
<td>Accumulation rate x Dry bulk density</td>
</tr>
<tr>
<td>Pollen concentration wet weight</td>
<td>10⁶ grains g⁻¹</td>
<td>[(Lyco added x Pollen counted / Lyco counted) / sediment wet weight] x 10⁶</td>
</tr>
<tr>
<td>Pollen concentration dry weight</td>
<td>10⁶ grains g⁻¹</td>
<td>(Pollen conc wet weight x 100) / %DW</td>
</tr>
<tr>
<td>Pollen accumulation rate</td>
<td>10⁶ grains cm⁻² yr⁻¹</td>
<td>Pollen conc dry weight x Dry mass accumulation rate</td>
</tr>
</tbody>
</table>

Squares (CONISS) cluster analysis was carried out for each pollen sequence, using the Edwards and Cavalli-Sforza chord distance as a dissimilarity measure, which is usually the most suitable dissimilarity coefficient for pollen data (Grimm, 1987). CONISS is an agglomerative method that creates stratigraphically constrained clusters based on the similarity of adjacent samples or groups of samples, and produces a hierarchical dendrogram that can be displayed alongside the pollen data. Taxa that did not attain a level of ≥ 1% abundance in at least one sample were excluded from the CONISS analysis (Birks and Gordon, 1985). To determine the optimum number of pollen zones, a broken-stick model was applied (Bennett, 1996), whereby the CONISS results were compared with the expected results for a dataset of random samples. If the reduction in variance for a particular zone (as determined by the CONISS output) was greater than the proportion expected from the broken-stick model (which is based on a random dataset), then the zonation was considered ‘significant’ (Bennett, 1996: 161).

Pollen data were further analysed by multivariate ordination techniques, which provide a way to analyse floristic composition without stratigraphical constraints, i.e. based solely on the distribution of taxa within each sample. Multivariate ordination techniques arrange species along hypothetical axes, or gradients, and can thus help to define major patterns in the dataset that may not be particularly apparent from other stratigraphically-based
methods, such as the similarity of non-contiguous samples. Such techniques aim to find (usually hypothetical) axes of the 'greatest variability in the community composition for a set of samples and to visualise (using an ordination diagram) the similarity structure for the samples and species' (Leps and Šmilauer, 2003: 26). Several different multivariate ordination techniques can be applied to palaeoenvironmental datasets. An unconstrained ordination is one that does not include any measured environmental variables; instead, unconstrained ordination finds the latent gradients (i.e. ordination axes) that best represent the predictors for the values of all the species (Leps and Šmilauer, 2003). The model of species responses to the latent gradients must be specified – i.e. whether the species are showing a linear or unimodal response to the hypothetical variables. The first gradient (first ordination axis) length of Detrended Correspondence Analysis (DCA), a technique that estimates heterogeneity in community composition, can be used to determine whether linear or unimodal methods are most suitable for the dataset in question (Hill and Gauch, 1980; ter Braak and Prentice, 1988; Leps and Šmilauer, 2003).

An unconstrained ordination was applied to the pollen data from both sites using the programme CANOCO v. 4.5 (ter Braak and Šmilauer, 2002). Only species that attained a minimum of 1% abundance (total sum) in at least one sample were included in the ordination. The main patterns of floristic variation were assessed initially by DCA, and the gradient length of the first DCA axis was used to establish whether the responses of species were predominantly linear or unimodal. The appropriate ordination was then carried out – for linear responses, this involved Principal Components Analysis (PCA), with the application of Aitchison's (1990) log-ratio method whereby species data are log-transformed with centring by species and by samples.

4.4.2 Charcoal analysis

Charcoal is produced by the incomplete combustion of organic material, and is often incorporated into lake and swamp sediments (Patterson et al., 1987). Quantification of sedimentary charcoal can provide a means of reconstructing fire histories over local to regional scales (Patterson et al., 1987; Figueiral and Mosbrugger, 2000; Whitlock and Larsen, 2001; Conedera et al., 2009; Mooney and Tinner, 2011). Variations in the abundance of charcoal in a sediment sequence are assumed to reflect fire history, although
variation of taphonomic processes – the processes that transport and deliver charcoal to the site of deposition – may also be a factor (Whitlock and Larsen, 2001). The original amount of charcoal produced is modified by dispersal, deposition, preservation, sampling and counting methods (Patterson et al., 1987). Fragmentation of charcoal takes place due to differential shrinking as dehydration progresses, resulting in the production of charcoal fragments ranging in size from sub-microscopic to several cubic centimetres; the brittle nature of charcoal, however, means that larger fragments can easily be broken into smaller (Clark, 1984). Charcoal particles may be transported by wind, water, or erosional processes to the site of deposition (Scott et al., 2000). Charcoal emitted to the atmosphere will be dispersed according to the gravimetical laws that govern all small particles, such that large heavy particles and those with a high ratio of volume to surface area tend to move shorter distances (Patterson et al., 1987; Thevenon et al., 2003). Small particles carried away from a fire in smoke may travel great distances (Clark, 1984). However, the transport of charcoal particles from a fire to a sedimentary basin is dependent on several factors, including the size of particles, the intensity of the fire and the weather conditions (Mooney and Tinner, 2011). Reconstructing the intensity, size, distance and direction of fires from the site of charcoal deposition can therefore be problematic. Nonetheless, sedimentary charcoal can be a sensitive indicator of palaeoenvironmental changes, and charcoal abundance is often related to palaeoclimate, past volcanic activity and/or past vegetation change (including the influence of anthropogenic activity on vegetation) (e.g. Patterson et al., 1987; Blackford, 2000; Figueiral and Mosbrugger, 2000).

Macroscopic charcoal (macrocharcoal) refers to charcoal fragments larger than about 150 μm in length, and is thought to represent stand- to local-scale fire history, depending on sampling site area (Whitlock and Millspaugh, 1996). This is based on the fact that macrocharcoal is unlikely to be transported far in the atmosphere, and therefore is likely to constitute a reliable proxy of local fires (Carcaillet, 2007). Samples from both the lake and swamp sites were analysed for macroscopic charcoal by researchers at the School of Biological, Earth and Environmental Sciences, University of New South Wales, Australia. The method involved bleaching, sieving and photographing sediment samples, followed by the calculation of charcoal area using image analysis software (see Mooney and Tinner
Charcoal influx was also calculated by multiplying charcoal concentration by the sedimentation rate (as calculated by the age model), and expressed as area of charcoal (mm$^2$) per area of sediment (cm$^2$) per year.

Microscopic charcoal (microcharcoal) refers to charcoal fragments smaller than about 150 μm in length. As particles of this size are usually well-transported by wind (Carcaillet, 2007), microcharcoal is considered a proxy of fire at all scales up to and including a regional or extra-regional source area (Clark, 1988). Measurement of microcharcoal on pollen slides has been a common quantification method since the pioneering work of Iverson (1941), and several different measurement techniques are available, based on expressing the concentration of charcoal as a count or as an area (Whitlock and Larsen, 2001). Microcharcoal was quantified on the slides that had been prepared for pollen analysis from the Mt. Muhavura crater lake core sequence, which had included the addition of Lycopodium spores. Samples from the Mt. Gahinga crater swamp core sequence contained too low a concentration of microcharcoal to enable accurate quantification by pollen-slide methods. Charcoal is here regarded as jet black, uniformly opaque, angular particles with a long axis greater than 2.5 μm (Whitlock and Larsen, 2001). Two techniques of microcharcoal quantification were employed (both at x 400 magnification on a Leica DMLS light microscope): point-counting and size-classing.

The point-counting technique (Clark, 1982) was used to calculate the surface area of microcharcoal per unit volume of sediment (i.e. cm$^2$ cm$^{-3}$) for each sample. The number of points (intersections on an eyepiece grid) coinciding with fragments of charcoal were counted in 500 fields of view, spaced along 2 mm transects. The number of Lycopodium spores encountered along the transects was also recorded. Based on the theory of point-counting – which indicates that the probability of any point applied at random on a plane surface intercepting a particular phase (in this case, charcoal) will be the ratio of the area of the phase to the total area of the plane (Clark, 1982) – the area of microcharcoal in 1 cm$^3$ of sediment was calculated.

Size-classing of microcharcoal followed an adaptation of the method devised by Waddington (1969), whereby charcoal fragments were placed into size-classes based on
their maximum dimensions: 2.5 to 12.5 μm; 12.5 to 25 μm; 25 to 45 μm; 45 to 63 μm; 63 to 100 μm; and 100 to 140 μm (fragments < 2.5 μm were deemed too small to allow recognition as charcoal to an acceptable level of certainty). Based on the average area of fragments in each size-class, the total area of microcharcoal in 1 cm³ of sediment was calculated. The two estimates of microcharcoal area (point-counting and size-classing) could thus be compared.

4.4.3 Tephra analysis

The term ‘tephra’ refers to a wide range of airborne pyroclastic material ejected during a volcanic eruption (Turney and Lowe, 2001). The ash component has a grain size of less than 2 mm. Ash is often erupted very high into the atmosphere, and can therefore be dispersed widely following an eruption, and become incorporated into sediments accumulating across a range of different sedimentary environments. Generally both horizon thickness and mean grain size of tephra decline with increasing distance from eruptive sources (Turney and Lowe, 2001). However, if eruptions are not explosive, then tephra deposits may not occur, even at proximal sites.

Tephra deposits with characteristics by which they can be uniquely identified have significant potential for correlation and dating of sediment records. Regional tephrochronological frameworks for the late Quaternary have been (and continue to be) extensively developed for many parts of the world, particularly Europe (e.g. Pilcher and Hall, 1992; Davies et al., 2002; Turney et al., 2006; Lowe et al., 2007; Lowe et al., 2008; Davies et al., 2010; Blockley et al., in press; Lane et al., in press). This has been aided by the development of techniques to analyse cryptotephra horizons – i.e. tephra horizons that cannot be seen with the naked eye (e.g. Turney et al., 1997; Blockley et al., 2005; Gehrels et al., 2008). In this thesis, the term ‘tephra’ refers primarily to cryptotephra, unless otherwise specified. In Africa, a comprehensive tephrochronological framework has yet to be developed: the European framework is currently being extended to parts of northern Africa (Lane et al., 2011), but tephra horizons in late Quaternary sediments from eastern Africa cannot yet be used as chronological markers, although significant potential exists to develop a framework in the region (Pyle, 1999; Olago et al., 2000). However, even without a chronological framework, tephra horizons may be an important indicator of
palaeoenvironmental change – and, in the case of proximal eruptions, may be related to aquatic ecosystem perturbation (Barker et al., 2000) or to fire history (Ogden et al., 1998). Late Quaternary tephra horizons at two sites in southwest Uganda have been linked to possible eruptive activity in the Virunga volcanoes, but identification of a single source area proved difficult (Taylor, 1988).

Given the potential of tephra horizons to act as correlative markers between sites, in addition to potential (albeit tentative) links between tephra deposition and aquatic ecosystem and/or vegetation change, the cores from the Virunga volcanoes obtained in this study provided suitable test cases for tephra analysis. No visible tephra horizons were present in the sediment cores, but parts of the cores with significant fluctuations in the physical proxy measurements (particularly magnetic susceptibility) were identified as potentially containing cryptotephras. Preliminary assessment of tephra content was carried out following the procedure of Pilcher and Hall (1992). Contiguous samples over 5 cm were taken, at intervals of ca. 15 cm, from throughout the cores from the crater lake on Mt. Muhavura. Samples were oven dried and then ignited at 600°C for four hours, and the resulting ash was rinsed with 10% HCl (to remove the soluble inorganic fraction), washed and mounted onto slides with Naphrax® (refractive index 1.73). Tephra was detected by optical microscopy using a polarising microscope, and identified by its colour, vesicularity and morphology (Pilcher and Hall, 1992; Turney and Lowe, 2001; Enache and Cumming, 2006). Tephra is usually colourless and is recognised under an optical microscope by its vesicular nature and lack of crystal structure. Fragments of vesicular glass with enclosed gas bubbles can be taken as diagnostic.

To confirm the preliminary identification of tephra shards in the sediment cores, fifteen samples from the Muhavura cores and five samples from the Gahinga cores were analysed by researchers at Swansea University, using a density separation technique (modified from Blockley et al., 2005). Samples were freeze-dried and 0.5 g (dry weight) of sediment was prepared for tephra analysis. The samples were ashed due to their high organic content, HCl was used to remove any carbonates, and the remaining material floated at densities of 2.5 and 2.3 g cm⁻³ using sodium polytungstate. Counts of tephra, using optical microscopy, were carried out on these samples.
4.4.4 Diatom analysis

Diatoms (class: Bacillariophyceae) are unicellular photosynthetic algae with a characteristic siliceous outer shell. They occur in almost all aquatic environments, and constitute a major component of algal assemblages (Battarbee et al., 2001). Their uptake and deposition of silica leads to rigidity of the cell wall, which consists of two intricately sculptured valves, linked together with girdle bands. Diatom shape varies widely and, together with the characteristics of the highly patterned cell wall, is used for identification purposes (Jones, 2007). The abundance, ubiquity, good preservation in sediments, ability to respond rapidly to environmental change, and often well-known ecological preferences of diatoms has led to their extensive use as palaeolimnological proxies (Stoermer and Smol, 1999; Battarbee et al., 2001). Past diatom communities can be reconstructed from lacustrine sedimentary records, thus providing an insight into changing limnological conditions – including lake level change – and catchment dynamics (e.g. Laird et al., 1998; Barker et al., 2000; Barker et al., 2002; Legesse et al., 2002). Diatoms rarely have very discrete habitats, as species from different communities often overlap. In lakes, the distinction between planktonic and benthic diatoms is often not absolute: some planktonic diatoms spend some of their life-cycle resting on the sediment (meroplanktonic), while some species which have their true habitat in the benthos can be found resuspended in the water column (tychoplanktonic). Benthic diatoms are often very diverse and can live either free on or in the sediments (epipelagic), or attached to the substratum, e.g., rocks (epilithic) and sand grains (epipsammic) or to plants (epiphytic) (Jones, 2007).

In this study, 30 samples (0.1 g wet weight) from the sediments extracted from the crater lake on Mt. Muhavura were prepared for diatom analysis following standard procedures (Battarbee et al., 2001), involving removal of organic material with 30% H₂O₂ and addition of a few drops of 10% HCl to remove any carbonates. To enable calculation of diatom concentrations, a known weight of DVB (divinylbenzene) microspheres was added to the cleaned samples (Battarbee and Kneen, 1982). Suspensions were diluted to a suitable concentration and mounted on slides with Naphrax®, for counting and identification. Diatoms were counted along 1 mm transects at x 1000 magnification under an oil-immersion objective and phase contrast on a Meiji ML.5000 Phase Contrast light
microscope. A minimum of 400 valves was counted in each sample to ensure that a representative count was obtained (Battarbee et al., 2001). Diatoms were identified according to a range of published papers and books, primarily Gasse (1986), Krammer and Lange-Bertalot (1986-1991), Krammer (1992), Lange-Bertalot and Moser (1994), Cocquyt (1998) and Cocquyt and Jahn (2007).

Diatoms were expressed both as relative percentages and as concentrations (in $10^6$ valves per gram dry weight). Total diatom concentration and accumulation rate were calculated, following the same method as that for pollen calculations (Table 4.1). Zonation of diatom data was carried out using CONISS (Grimm, 1987), and the significant number of diatom zones was evaluated using a broken-stick model (Bennett, 1996) (see Section 4.4.1). Multivariate ordination techniques (using diatom percentage relative abundances) were carried out using CANOCO v.4.5 (ter Braak and Šmilauer, 2002), following the same methods as those outlined for the pollen data (Section 4.4.1). Diatom taxa that attained a relative abundance of $\geq 1\%$ in at least one sample were included in the numerical analyses.

4.4.5 Organic isotope analysis

Analysis of the various constituents of organic matter in lake sediments can elucidate important environmental changes in the lake and its catchment. Geochemical analysis of lake sediment organic matter can yield information about the source and amount of organic material entering a lake system (Meyers and Teranes, 2001). The most commonly measured parameters of lake sediment organic matter are percentage total organic carbon (%TOC), carbon/nitrogen (C/N) ratios, and carbon isotope ratios ($\delta^{13}C$), although analysis of percentage organic nitrogen (%TON) and nitrogen isotope ratios ($\delta^{15}N$) is becoming increasingly common (Meyers and Teranes, 2001; Talbot, 2001). %TOC provides an indication of organic matter abundance in sediments. Typical organic matter contains approximately 50% carbon, so %LOI values are usually equivalent to about twice the %TOC values (Meyers and Teranes, 2001). C/N ratios provide a useful way to distinguish between algal and land-plant sources of sedimentary organic matter (Meyers, 1994). Organic matter from phytoplankton has C/N ratios that are commonly between 4 and 10, whereas vascular land plants, which are cellulose-rich and protein-poor, create organic matter that usually has C/N ratios of 20 and greater (Meyers and Teranes, 2001). In
sediments with 1% or more TOC, C/N ratios are normally reliable (Meyers, 2003). \( \delta^{13}C \) values may provide information on organic matter sources, past productivity rates and/or changes in nutrient availability, although distinguishing the factors to which \( \delta^{13}C \) is responding is sometimes problematic (Meyers and Teranes, 2001). Further information may be provided by measuring %TON and \( \delta^{15}N \) contents of lake sediment organic matter. Nitrogen is a key nutrient in lacustrine systems, and is viewed as one of the nutrients that ultimately limits organic productivity (Talbot, 2001). \( \delta^{15}N \) of lacustrine organic matter (\( \delta^{15}N_{OM} \)) can have a wide range of values, largely depending on the isotopic composition of the nitrogen source and the nitrogen metabolism of the dominant taxa (Wolfe et al., 1999); changes in nitrogen cycling within a lake are also likely to cause some significant isotopic effects (Talbot and Laerdal, 2000). While diagenetic alteration cannot be ruled out as potentially affecting the composition of sediment organic matter, particularly in the early stages of deposition, source and palaeoenvironmental information may nonetheless remain preserved in its elemental and isotopic composition (Meyers, 1994; Hodell and Schelske, 1998). A multi-proxy geochemical archive may thus provide important insights into changing biota in a lake and its surrounding area (Meyers and Ishiwatari, 1993).

Geochemical analysis of sediment organic matter was carried out on samples from the core sequence from the crater lake on Mt. Muhavura. A total of 68 samples were pre-treated with 5% HCl to remove any carbonates, washed through filter paper, dried at 50°C, and ground to a fine powder. Percentage carbon and nitrogen (to calculate C/N ratios), \( ^{13}C/^{12}C \) and \( ^{15}N/^{14}N \) analyses were carried out on a Thermo Delta plus Continuous Flow Isotope Ratio Mass Spectrometer (CF-IRMS) coupled to a CE Instruments 1112 Flash Elemental Analyser in the Geochemistry Laboratory of the Department of Geology, Trinity College Dublin. \( \delta^{13}C \) values were calculated to the VPDB scale using a within-run laboratory standard.
Chapter 5 Results I: Core chronologies and physical proxies

This chapter presents the results of the field- and laboratory-based analyses of sediment cores from crater sites on Mt. Gahinga and Mt. Muhavura. In particular, the chapter focuses on establishing chronological control for sedimentary records obtained from both study sites. The age models produced from the radiometric dates, based on Bayesian techniques, are presented and discussed. Correlation of the cores from the crater lake on Mt. Muhavura is also described and discussed. Results from the crater swamp on Mt. Gahinga are presented first, including details of the sediment core extraction, description, chronological control, age model production, and physical proxy results, followed by those from the crater lake on Mt. Muhavura. For the latter site, physical proxy data are presented before development of the age models is described, as these results are used to help inform the best option for core correlation and age determination.

5.1 Mt. Gahinga crater swamp

5.1.1 Sediment core extraction

Fig. 5.1 summarises variations in sediment depth across the crater swamp on Mt. Gahinga. Measurements were taken every 10 m along a transect running in a NE–SW direction (between compass bearings 50° and 230°). At the centre of the transect, four test locations (at 30 m, 40 m, 50 m and 60 m) extended to a sediment depth greater than 9 m (the length of the depth-testing rods). The coring site was selected as the centre of these test locations (Fig. 5.1) (GPS co-ordinates: 1° 23.157' S; 29° 38.722' E). Cores extending from 0.5 m to a maximum sediment depth of 9.74 m were obtained. Based on the morphology of the crater as indicated by the depth transect, the maximum depth of peat deposition in the centre of the crater is estimated as ca. 14 m, but equipment restrictions meant that only ca. 10 m of sediment could be collected.

Sediment stratigraphy was relatively homogenous throughout the sequence (Fig. 5.1), consisting of highly organic herbaceous peat with an abundance of moss and other herbaceous plant fragments. A slight increase in minerogenic content was apparent during sub-sampling of the sediment cores at depths of 700–702 cm and 600–616 cm (Fig. 5.1).
Fig 5.1: Variations in sediment depth across Mt. Gahinga crater swamp, showing the location of the coring site and sediment stratigraphy.
5.1.2 Radiocarbon dating

The results of radiocarbon analysis of the nine macrofossil samples from the Mt. Gahinga crater swamp cores are summarised in Table 5.1, including their conventional radiocarbon ages and estimated errors (at 2σ), calibrated age range (using the IntCal09 calibration curve), and δ¹³C values. Photographs of three of the samples are shown in Plate 5.1. The calibrated age range is taken as the oldest and youngest ages from the calibrated date, and do not represent a normal distribution. The series of plots in Fig. 5.2 show the details of the calibration process, and highlight the complexity of the multi-modal probability distributions produced by calibration, which arise due to the non-linearity of the calibration curve. Some calibrated dates show probability distributions that are near-normal (e.g. GAH10 940-941 cm, Fig. 5.2i), but others show marked split distributions (e.g. GAH2 150-151cm, Fig. 5.2b) or very wide distributions (e.g. GAH8 746-747 cm, Fig. 5.2h). In such cases, taking the mid-point of the calibrated age would not be an accurate age approximation, as this would ignore the information contained in the multi-modal probability distribution.

The age–depth profile of the calibrated age ranges for the dated samples is plotted in Fig. 5.3, and shows an increase in age with depth, with no apparent age reversals or sedimentary hiatuses. The calibrated age ranges are mostly relatively narrow, although several samples show split distributions (cf. Fig. 5.2). The sedimentation rate appears to increase slightly from about 400 cm upwards. Overall the age–depth plot suggests continuous sedimentation throughout the core sequence, which further confirms the suitability of this site for palaeoenvironmental reconstruction.

The base of the sequence appears to date to between ca. 8000 and 7500 cal yrs BP. The top of the core sequence is at a depth of 50 cm, and the topmost dated sample (from a depth of 120-121 cm) has a 63-69% chance of dating within the range 321 to 282 cal yrs BP. The ¹⁴C dates thus imply that the sequence extends from ca. 8000 cal yrs BP to ca. 200 cal yrs BP. This age–depth relationship is analysed further in the next section using a Bayesian age model.
Plate 5.1: Photos of three of the plant macrofossil samples (moss fragments) from the Mt. Gahinga crater swamp sediments from which radiocarbon dates were obtained. The scale bar applies to all three images.

Table 5.1: Results of radiocarbon dating of macrofossil samples from the Mt. Gahinga core sequence. The sample from GAH10 940-941cm was analysed at Beta Analytic Inc., Florida. All other samples were analysed at the CHRONO Centre, Queen’s University, Belfast. All radiocarbon ages were calibrated to calendar years using the IntCal09 calibration curve (Reimer et al., 2009). Cal yrs BP is years before 1950.

<table>
<thead>
<tr>
<th>Core</th>
<th>Depth</th>
<th>Lab ID number</th>
<th>AMS δ¹³C (%)</th>
<th>Conventional radiocarbon age (¹⁴C yrs)</th>
<th>Calibrated age range (2σ) (cal yrs BP)</th>
</tr>
</thead>
<tbody>
<tr>
<td>GAH2</td>
<td>120-121cm</td>
<td>UBA-13563</td>
<td>-25.7</td>
<td>265 ± 23</td>
<td>428 152</td>
</tr>
<tr>
<td>GAH2</td>
<td>150-151cm</td>
<td>UBA-12322</td>
<td>-22.4</td>
<td>350 ± 25</td>
<td>492 316</td>
</tr>
<tr>
<td>GAH3</td>
<td>219-220cm</td>
<td>UBA-13564</td>
<td>-30.1</td>
<td>844 ± 23</td>
<td>790 695</td>
</tr>
<tr>
<td>GAH3</td>
<td>270-272cm</td>
<td>UBA-13565</td>
<td>-26.8</td>
<td>943 ± 24</td>
<td>922 795</td>
</tr>
<tr>
<td>GAH4</td>
<td>321-322cm</td>
<td>UBA-13566</td>
<td>-27.0</td>
<td>1397 ± 24</td>
<td>1344 1285</td>
</tr>
<tr>
<td>GAH4</td>
<td>374-375cm</td>
<td>UBA-12324</td>
<td>-28.6</td>
<td>1660 ± 22</td>
<td>1686 1521</td>
</tr>
<tr>
<td>GAH6</td>
<td>550-551cm</td>
<td>UBA-12325</td>
<td>-21.3</td>
<td>3414 ± 25</td>
<td>3811 3583</td>
</tr>
<tr>
<td>GAH8</td>
<td>746-747cm</td>
<td>UBA-12326</td>
<td>-23.1</td>
<td>5270 ± 34</td>
<td>6180 5937</td>
</tr>
<tr>
<td>GAH10</td>
<td>940-941cm</td>
<td>Beta-250096</td>
<td>-20.2</td>
<td>6800 ± 40</td>
<td>7690 7580</td>
</tr>
</tbody>
</table>
Fig. 5.2: 2σ calibration of the nine radiocarbon dates from the Mt. Gahinga core sequence, showing the resultant probability distributions. All dates were calibrated against IntCal09 using OxCal v. 4.1.7.
5.1.3 Age model production

All radiocarbon dates from the Mt. Gahinga core sequence were input into a P-sequence (i.e. Poisson distribution) model in OxCal v. 4.1.7. The top and bottom depths of the sediment sequence were also input as boundaries, which are arbitrary dates placed at the beginning or end of a dated sequence. Depth values for boundaries are specified in order to extrapolate the model beyond the dated points in the sequence (Bronk Ramsey, 2008). Furthermore, an age range of 1800 to 2008 AD was assigned to the top boundary (50 cm). This boundary was set as wide as possible, so as to not overly restrict the model, but in order to sufficiently constrain this age between the youngest date (152 cal yrs BP) in the calibrated age range of the topmost radiocarbon date (120-121 cm) and the present day, to prevent the model from dating the top of the sequence to a date in the future. No age range was input to constrain the lowermost boundary (the base of the sequence, i.e. 974 cm).

Results from the model output are summarised in Table 5.2, and the age–depth model is plotted in Fig. 5.4. An overall agreement index ($A_{\text{overall}}$) of 64% is given for a P-sequence model with $k = 0.6$. A high agreement index for a date occurs when the modelled date
constrained stratigraphically by the model) is similar to the unmodelled date (i.e. calibrated, but not constrained stratigraphically). A P-sequence model with $k = 0.7$ yielded an $A_{\text{overall}}$ of 57.7%. As 60% is taken as the threshold for acceptable agreement index (Bronk Ramsey, 2008), the model with $k = 0.6$ is taken as the final model.

Comparisons of some of the modelled dates with their corresponding unmodelled, calibrated distributions are shown in the boxed graphs (a-e) in Fig. 5.4. In some cases (e.g. Fig. 5.4c and 5.4e), the modelled and unmodelled distributions are almost identical, yielding an agreement index of ca. 100%. Fig. 5.4d (GAH8-746cm) highlights a case where the modelled date only overlaps with a portion of the unmodelled date, but the agreement index is still high (91.7%). GAH4-374cm, on the other hand, has a low agreement index (23.9%) because the modelled date overlaps with a part of the unmodelled date that has a relatively low probability within the calibrated probability distribution (Fig. 5.4b). This is the only date with an agreement index falling below 60%. Overall, the P-sequence model provides an improved age-depth relationship compared with the raw

Table 5.2: OxCal model output for a P-sequence model for the Mt. Gahinga radiocarbon dates. Overall agreement index ($A_{\text{overall}}$) is 64% ($k = 0.6$). The agreement index for each modelled date is also shown (A), which only falls below 60% for one date (374-375 cm).
Fig. 5.4: Age–depth model for the core sequence from the crater swamp on Mt. Gahinga, using a P-sequence model with $k = 0.6$. Details of five of the modelled dates are shown as examples of how the modelled distributions compare with the unmodelled, calibrated distributions.
calibrated radiocarbon dates in Fig. 5.3 for two reasons. Firstly, because the model utilises the relative sediment depth of the calibrated radiocarbon dates to analyse which portion of the multi-modal probability distribution is most likely, while allowing for fluctuations in sedimentation rate. Secondly, because an improved estimation of the sediment age between samples, which is narrower than that derived from using the entire calibrated ranges of the dates, is provided.

The model assigns a bottom date to the sequence (at 974 cm) of between 8085 and 7786 cal yrs BP, indicating that the sediment sequence dates back to the early Holocene. The modelled date for the top of the sequence, however, is more uncertain, as it assigns a date (at 50 cm depth) of -7 to -59 cal yrs BP (i.e. 1943 to 2008 AD). Without further independent dating information on the topmost sediments, and without a core sequence extending continuously downwards from the swamp surface, the dating of the uppermost 120 cm of sediment (i.e. above the topmost radiocarbon date) must remain somewhat tentative.

The age model presented in Fig. 5.4 was used as the basis for developing a secondary y-axis (cal yrs BP) to display alongside sediment depth in the plots of physical, biological and chemical proxy measurements shown from this point onwards (see e.g. Fig. 5.5).

5.1.4 Physical proxy measurements

Fig. 5.5 shows the results of all physical proxy measurements (%LOI, % dry weight, wet density, and magnetic susceptibility) for the cores from the Mt. Gahinga crater swamp. The core sequence consists of highly organic, homogenous sediment, with only two major exceptions, both consisting of a short-lived rise in minerogenic content. %LOI values are very high – greater than 90% for most samples – indicating very high organic content throughout the sequence, which is not uncommon for swamp sediments. Between ca. 250 cm and 175 cm, %LOI values drop slightly, to about 80%. The most distinct features of the %LOI measurements are the abrupt decreases to 50% at 700 cm (ca. 5100 cal yrs BP) and to 28% at 616 cm (ca. 4100 cal yrs BP). Distinct peaks in % dry weight and wet density also occur at both these depths. The magnetic susceptibility measurements mostly registered as values close to zero, due to the high organic content in the sediment cores. Two magnetic susceptibility peaks occur: a small peak at 700 cm and a major peak at
616 cm, both of which represent a major drop in sediment organic content, and increase in coarser-grained, minerogenic content. This was also observed visually in the sediment sequence, whereby denser, sandier material occurred in these parts of the sequence.

Overall, the results from the Mt. Gahinga crater site thus show that the 974-cm core sequence extends back to between ca. 8000 and 7800 cal yrs BP, although uncertainty remains about the age of the uppermost section of the core sequence. The physical proxy measurements for the Mt. Gahinga core sequence showed two significant short-lived fluctuations in sediment content at 700 cm and 616 cm depth (dating to ca. 5100 and 4100 cal yrs BP, respectively), consisting of a marked increase in minerogenic, coarse-grained material. There is a further, but less marked, drop in organic content between ca. 250 and 150 cm (ca. 850 to 430 cal yrs BP).
5.2 Mt. Muhavura crater lake

5.2.1 Sediment core extraction

Transect lines across the crater lake on Mt. Muhavura used to guide coring were run in N-S and E-W directions (Fig. 5.6). A maximum water depth of 1.3 m, and maximum sediment depth of 3.41 m, were recorded along the transects. The area of deepest sediment accumulation was near the centre of the lake. The lake has a Secchi disc depth of 0.49 m (measured in July 2008 at a water depth of 0.9 m, at 11.30 am under a cloudy sky). The lake's water is mildly acidic (pH 6.1) and well-oxygenated (surface dissolved oxygen 8.6 mg/l), and has a surface water temperature of 9.1°C.

Two sites (GPS co-ordinates: 1° 22.97' S; 29° 40.67' E) were chosen from which to extract cores using the Livingstone piston corer: site MUH1, almost in the centre of the lake, and site MUH2, slightly away from the centre of the lake, and 5 m from MUH1 (Fig. 5.6). The cores extracted reached a maximum sediment depth of 277 cm at site MUH1, and 292 cm at site MUH2. As casing to mark the coring hole was not used, overlapping cores were taken. Each overlapping core extracted at site MUH1 and site MUH2 was assigned a sequential number (see Table 5.3). Each core was extracted over a sediment depth of one metre, with the exception of core MUH2-4 (the base of the MUH2 sequence), which only penetrated the sediment between 235 cm and 292 cm (i.e. a 57 cm core). Most cores required a small correction for core shortening. Two cores (MUH1-1 and MUH2-1) required a correction for core expansion, which was mostly likely due to the high water content in these cores. Core MUH1-1 and, to a lesser extent, core MUH2-2, displayed possible signs of sediment slumping or mixing, due to the high retention of water in the coring barrel and expansion upon extrusion, which meant they each had to be extruded into two separate pieces of guttering.

Logs of the overlapping cores for sites MUH1 and MUH2 are shown in Fig. 5.7, indicating the depths (and relative depths) at which each core was extracted, and showing the Troels-Smith descriptions of the sediment (Aaby and Berglund, 1986). The sediment was mostly homogenous, organic-rich lake mud. Water content was observed to be generally high in
Fig. 5.6: Schematic drawing of Mt. Muhavura crater lake transects and measured depths, and showing the location of coring sites in this study.
Table 5.3: List of cores extracted at the two coring sites in the Mt. Muhavura crater lake. All cores except core MUH2-4 required a small correction for core shortening or (in two cases), core expansion.

<table>
<thead>
<tr>
<th>Core name</th>
<th>Measured depth at extraction</th>
<th>Length of core before correction</th>
</tr>
</thead>
<tbody>
<tr>
<td>MUH1-1</td>
<td>71 to 171 cm</td>
<td>127 cm *</td>
</tr>
<tr>
<td>MUH1-2</td>
<td>100 to 200 cm</td>
<td>91 cm</td>
</tr>
<tr>
<td>MUH1-3</td>
<td>177 to 277 cm</td>
<td>86 cm</td>
</tr>
<tr>
<td>MUH2-1</td>
<td>50 to 150 cm</td>
<td>106 cm *</td>
</tr>
<tr>
<td>MUH2-2</td>
<td>125 to 225 cm</td>
<td>98 cm</td>
</tr>
<tr>
<td>MUH2-3</td>
<td>168 to 268 cm</td>
<td>90 cm</td>
</tr>
<tr>
<td>MUH2-4</td>
<td>235 to 292 cm</td>
<td>57 cm</td>
</tr>
</tbody>
</table>

*Due to the length of cores MUH1-1 and MUH2-1 (due primarily to water expansion), these cores were both extruded into two separate pieces of guttering.

the upper parts of the sequence, and reduced with depth. Herbaceous fragments (particularly mosses) occurred throughout the cores. A number of short-lived marked changes in sediment composition were observed, with increased silt and sand content. The most marked occurrence of this, at MUH2-4 256-257 cm and MUH2-3 253-254 cm, also showed a distinctive change in sediment colour (see photographs in Fig. 5.8). A similar change in the sediment stratigraphy occurred at MUH1-3 227-228 cm. The other sections of the cores with increased silt and sand content (MUH1-2 135-138 cm; MUH2-2 130-131 cm and MUH2-1 80-86 cm) did not have as clear a visual change in sediment colour, but the change in the dominant grain size of the sediment matrix was apparent upon closer inspection and description.

A 37 cm core (MUH4) of the uppermost sediment, which included the sediment-water interface, was successfully collected using a Renberg corer. The site for extracting core MUH4 was close to the MUH1 and MUH2 sites, but not immediately beside, as the surface sediment would have been disturbed (see Fig. 5.6 for coring location). Core MUH4 consisted of homogenous lake mud throughout the core, and no obvious changes in sediment stratigraphy were evident. The uppermost part of the core, close to the sediment-water interface, was extremely unconsolidated and had very high water content.
Fig. 5.7: Details of MUH1 and MUH2 core depths, as measured during core extraction. Sedimentary logs, based on Troels-Smith descriptions (Aaby and Berglund, 1986), are also shown. All cores have been corrected for shortening/expansion where necessary.
Fig. 5.8: Photographs of cores MUH2-3 and MUH2-4 taken prior to subsampling, highlighting details of the most noticeable visible change in the sediments that occurs in the MUH2 core sequence.
5.2.2 $^{210}\text{Pb}$ and $^{137}\text{Cs}$ radiometric dating

The results from $^{210}\text{Pb}$ and $^{137}\text{Cs}$ analysis of core MUH4 (37 cm length) are shown in Tables 5.4 and 5.5, and are plotted in Fig. 5.9 and 5.10. Total $^{210}\text{Pb}$ activity does not reach equilibrium with the supporting $^{226}\text{Ra}$ at the bottom of the core, indicating that background levels of $^{210}\text{Pb}$ activity were not reached. Unsupported $^{210}\text{Pb}$ activities, calculated by subtracting $^{226}\text{Ra}$ activity from total $^{210}\text{Pb}$ activity, show an overall steady decline with depth. There is a slight dip in unsupported $^{210}\text{Pb}$ activities from 18-24 cm, possibly suggesting slumping. The unsupported $^{210}\text{Pb}$ inventory of 20178 Bq m$^{-2}$ in the core corresponds to a mean $^{210}\text{Pb}$ supply rate of 673 Bq m$^{-2}$ yr$^{-1}$. This is significantly higher than the estimated atmospheric flux in the region, and indicates sediment focusing.

Fig. 5.9: Unsupported $^{210}\text{Pb}$ activity in core MUH4 from the crater lake on Mt. Muhavura.
Table 5.4: $^{210}$Pb concentrations in core MUH4 from the crater lake on Mt. Muhavura.

<table>
<thead>
<tr>
<th>Depth (cm)</th>
<th>Dry mass (g cm$^{-2}$)</th>
<th>$^{210}$Pb Total</th>
<th>Unsupported $^{210}$Pb</th>
<th>Supported $^{210}$Pb</th>
<th>Cumulative unsupported $^{210}$Pb</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td></td>
<td>$Bq$ kg$^{-1}$</td>
<td>$Bq$ kg$^{-1}$ ±</td>
<td>$Bq$ kg$^{-1}$ ±</td>
<td>$Bq$ m$^{2}$ ±</td>
</tr>
<tr>
<td>0.5</td>
<td>0.0197</td>
<td>2656.5</td>
<td>64.6</td>
<td>2586.6</td>
<td>69.9</td>
</tr>
<tr>
<td>2.75</td>
<td>0.1086</td>
<td>2483.0</td>
<td>74.3</td>
<td>2342.6</td>
<td>50.4</td>
</tr>
<tr>
<td>3.75</td>
<td>0.1851</td>
<td>2349.4</td>
<td>62.0</td>
<td>2305.0</td>
<td>44.4</td>
</tr>
<tr>
<td>5.5</td>
<td>0.3209</td>
<td>1740.3</td>
<td>31.5</td>
<td>1703.5</td>
<td>36.8</td>
</tr>
<tr>
<td>9.5</td>
<td>0.4853</td>
<td>1107.1</td>
<td>28.7</td>
<td>1073.0</td>
<td>34.1</td>
</tr>
<tr>
<td>11.5</td>
<td>0.5808</td>
<td>1152.7</td>
<td>16.6</td>
<td>1115.9</td>
<td>36.8</td>
</tr>
<tr>
<td>13.5</td>
<td>0.6862</td>
<td>766.8</td>
<td>20.2</td>
<td>729.6</td>
<td>37.2</td>
</tr>
<tr>
<td>15.5</td>
<td>0.823</td>
<td>964.8</td>
<td>22.3</td>
<td>930.5</td>
<td>34.2</td>
</tr>
<tr>
<td>17.5</td>
<td>0.9856</td>
<td>882.3</td>
<td>45.8</td>
<td>861.1</td>
<td>46.2</td>
</tr>
<tr>
<td>19.5</td>
<td>1.1481</td>
<td>510.4</td>
<td>24.4</td>
<td>474.3</td>
<td>24.7</td>
</tr>
<tr>
<td>23.5</td>
<td>1.4279</td>
<td>394.5</td>
<td>20.5</td>
<td>345.3</td>
<td>20.9</td>
</tr>
<tr>
<td>27.5</td>
<td>1.7739</td>
<td>489.6</td>
<td>50.8</td>
<td>448.6</td>
<td>51.7</td>
</tr>
<tr>
<td>31.5</td>
<td>2.0293</td>
<td>565.4</td>
<td>14.4</td>
<td>527.0</td>
<td>38.4</td>
</tr>
<tr>
<td>35.5</td>
<td>2.3429</td>
<td>361.3</td>
<td>20.2</td>
<td>312.8</td>
<td>20.6</td>
</tr>
</tbody>
</table>

Table 5.5: $^{137}$Cs concentrations in core MUH4 from the crater lake on Mt. Muhavura.

<table>
<thead>
<tr>
<th>Depth (cm)</th>
<th>$^{137}$Cs</th>
<th>$Bq$ Kg$^{-1}$ ±</th>
</tr>
</thead>
<tbody>
<tr>
<td>0.5</td>
<td>24.43</td>
<td>4.99</td>
</tr>
<tr>
<td>2.75</td>
<td>35.84</td>
<td>6.37</td>
</tr>
<tr>
<td>3.75</td>
<td>34.06</td>
<td>5.33</td>
</tr>
<tr>
<td>5.5</td>
<td>28.4</td>
<td>2.36</td>
</tr>
<tr>
<td>9.5</td>
<td>29.22</td>
<td>2.55</td>
</tr>
<tr>
<td>11.5</td>
<td>30.03</td>
<td>1.48</td>
</tr>
<tr>
<td>13.5</td>
<td>32.46</td>
<td>2.17</td>
</tr>
<tr>
<td>15.5</td>
<td>34.96</td>
<td>2.12</td>
</tr>
<tr>
<td>17.5</td>
<td>11.29</td>
<td>3.76</td>
</tr>
<tr>
<td>19.5</td>
<td>18.58</td>
<td>1.77</td>
</tr>
<tr>
<td>23.5</td>
<td>12.43</td>
<td>1.9</td>
</tr>
<tr>
<td>27.5</td>
<td>14.41</td>
<td>5.82</td>
</tr>
<tr>
<td>31.5</td>
<td>8.77</td>
<td>1.18</td>
</tr>
<tr>
<td>35.5</td>
<td>5.45</td>
<td>1.94</td>
</tr>
</tbody>
</table>
Fig. 5.10: $^{137}\text{Cs}$ concentration in core MUH4 from the crater lake on Mt. Muhavura. The $^{137}\text{Cs}$ concentration peak at 15.5 cm was used as a set reference point in the age-depth model.

Downcore variation in $^{137}\text{Cs}$ activity shows relatively poorly resolved peaks (Fig. 5.10): a lower peak at 15 cm and an upper peak at 2.5 cm. Relatively high values between these peaks may indicate mixing, although a decline overall in $^{210}\text{Pb}$ activity suggests any mixing has been minimal. The peak in $^{137}\text{Cs}$ concentration at 15.5 cm is most likely derived from the 1963 AD fallout maximum of the atmospheric testing of nuclear weapons. The sharp decline in $^{137}\text{Cs}$ in the deeper sediments just below the 15.5 cm peak may be due to slumping sedimentation, which diluted $^{137}\text{Cs}$ concentrations. The $^{210}\text{Pb}$-inferred age of the $^{137}\text{Cs}$ peak at 2.75 cm (2002 AD $\pm$ 6) is too young for it to be attributed to the 1986 AD Chernobyl nuclear accident, and may reflect mixing in the uppermost sediments.

A chronology for MUH4 was produced from the $^{210}\text{Pb}$ data using the constant rate of supply (CRS) model (Appleby and Oldfield, 1978). Use of the constant initial concentration (CIC) model was precluded by the non-monotonic variation in unsupported $^{210}\text{Pb}$ activity. The surface sediment age was taken as the date of core extraction, i.e. 2008 AD. The raw CRS model placed the 1963 AD depth at ca. 20 cm, but the final modelled dates were calculated by using the 1963 AD layer in the $^{137}\text{Cs}$ record as a reference point (Fig. 5.11).
Core MUH4 thus extends back to ca. 1900 AD, covering a period of just over 100 years in 37 cm of sediment. Calculated sedimentation rates based on this age–depth profile are also plotted in Fig. 5.11. The increased sedimentation rate around the 1950s could be due to sediment slumping; however, sedimentation rates are relatively stable over the last 50 years, with a mean at 0.019 g cm\(^{-2}\) yr\(^{-1}\). This relatively high sedimentation rate, combined with the high levels of \(^{210}\)Pb activity, suggests that core MUH4 reflects sediment focusing at the coring site during the last 100 years. If the core chronology produced for MUH4 is representative of other coring sites within the lake, then the sediment at a depth of 35.5 cm can be estimated as dating to 1904 AD ± 20.
5.2.3 Radiocarbon dating

The results of radiocarbon analysis of the five macrofossil samples extracted from the MUH1 core sequence and seven macrofossil samples from the MUH2 core sequence are shown in Table 5.6. The conventional radiocarbon age is given, along with the estimate of laboratory error (at 2σ), the calibrated age range, and the $^{13}$C measurements for each sample. Photographs of five of the samples from which radiocarbon dates were obtained are shown in Plate 5.2.

Details of the calibrations for each date are provided for the MUH1 dates in Fig. 5.12. As with the calibration details of dates for the Mt. Gahinga core sequence, the calibration plots highlight the complexity of the multi-modal distributions produced by calibration. In particular, MUH1-3 210-211 cm (Fig. 5.12d) shows a very split distribution (due to the fluctuation of the calibration curve), for which is it difficult to assign a best age estimate, and highlights a clear case where taking the mid-point of the calibrated age would not be an accurate age approximation. MUH1-3 181-182 cm has a very wide calibrated age range, due to the flattening of the calibration curve between ca. 2500 and 2400 $^{14}$C yrs BP (Reimer et al., 2009).

Plate 5.2: Photos of five of macrofossil samples (moss fragments) from the Mt. Muhavura crater lake sediments from which radiocarbon dates were obtained.
Table 5.6: Results of radiocarbon dating of macrofossil samples from cores MUH1 and MUH2 from the crater lake on Mt. Muhavura. Samples marked * were analysed at Beta Analytic Inc., Florida. All other samples were analysed at the 14CHRONO Centre, Queen’s University Belfast. All radiocarbon ages were calibrated using the IntCal09 calibration curve (Reimer et al., 2009). Cal yrs BP is years before 1950 AD.

<table>
<thead>
<tr>
<th>Core</th>
<th>Depth</th>
<th>Lab ID number</th>
<th>AMS δ13C (%)</th>
<th>Conventional radiocarbon age (14C yrs)</th>
<th>Calibrated age range (2σ) (cal yrs BP)</th>
</tr>
</thead>
<tbody>
<tr>
<td>MUH1-1</td>
<td>92-93 cm</td>
<td>UBA-13560</td>
<td>-21.0</td>
<td>325 ± 25</td>
<td>467 307</td>
</tr>
<tr>
<td>MUH1-2*</td>
<td>142-143cm</td>
<td>Beta-250097</td>
<td>-24.2</td>
<td>1900 ± 40</td>
<td>1928 1717</td>
</tr>
<tr>
<td>MUH1-3*</td>
<td>181-182cm</td>
<td>Beta-250098</td>
<td>-24.9</td>
<td>2450 ± 40</td>
<td>2705 2358</td>
</tr>
<tr>
<td>MUH1-3*</td>
<td>210-211cm</td>
<td>UBA-13561</td>
<td>-26.4</td>
<td>2271 ± 25</td>
<td>2348 2160</td>
</tr>
<tr>
<td>MUH1-3*</td>
<td>253-254cm</td>
<td>Beta-250099</td>
<td>-22.9</td>
<td>2610 ± 40</td>
<td>2843 2541</td>
</tr>
<tr>
<td>MUH2-1</td>
<td>65-66cm</td>
<td>UBA-13562</td>
<td>-26.2</td>
<td>548 ± 23</td>
<td>632 521</td>
</tr>
<tr>
<td>MUH2-1</td>
<td>107-108cm</td>
<td>UBA-12313</td>
<td>-23.6</td>
<td>1228 ± 23</td>
<td>1258 1070</td>
</tr>
<tr>
<td>MUH2-2</td>
<td>153-154cm</td>
<td>UBA-12314</td>
<td>-23.3</td>
<td>1495 ± 25</td>
<td>1478 1311</td>
</tr>
<tr>
<td>MUH2-2</td>
<td>185-186cm</td>
<td>UBA-12316</td>
<td>-23.6</td>
<td>1858 ± 25</td>
<td>1867 1723</td>
</tr>
<tr>
<td>MUH2-3</td>
<td>212-213cm</td>
<td>UBA-12317</td>
<td>-17.7</td>
<td>2087 ± 24</td>
<td>2125 1993</td>
</tr>
<tr>
<td>MUH2-3</td>
<td>242-243cm</td>
<td>UBA-12319</td>
<td>-25.6</td>
<td>2206 ± 23</td>
<td>2315 2150</td>
</tr>
<tr>
<td>MUH2-4</td>
<td>271-272cm</td>
<td>UBA-12320</td>
<td>-24.8</td>
<td>2463 ± 24</td>
<td>2705 2364</td>
</tr>
</tbody>
</table>

The 2σ calibrated age ranges for the MUH1 dates were plotted against depth (Fig. 5.13) to show the overall distribution of calibrated dates with depth for this core sequence. The MUH1 dates show one apparent age reversal, whereby the date obtained from MUH1-3 181-182 cm is older than would be expected based on the other dates. This may be due to sediment disturbance during coring, as this sample was taken from near the top of core MUH1-3 (top of core = 177 cm). Furthermore, core MUH1-3 showed the most core shortening (by an estimated 16 cm) of all extracted cores, and loss of sediment (or compression of sediment) cannot be discounted as potential factors in causing this anomalously old radiocarbon date. The other four dates show an increase in age with depth, but there is an apparent reduction in sedimentation rate towards the top of the sequence, as the topmost date (MUH1-1 92-93 cm) appears slightly younger than expected based on the trend suggested by the other ages. This could be a result of sediment mixing,
Fig. 5.12: 2σ calibration of the five radiocarbon dates from the MUH1 core sequence, showing the resultant probability distributions. All dates were calibrated against IntCal09 using OxCal v. 4.1.7.
Fig. 5.13: Age-depth profile for the MUH1 core sequence based on five radiocarbon dates, showing the calibrated age ranges. See Fig. 5.12 for the probability distributions of each date.

particularly in the top part of this core. The oldest date (MUH1-3 253-254 cm) lies between 3000 and 2500 cal yrs BP, although its probability distribution (Fig. 5.12e) shows that the highest probability age range (85.7%) occurs between 2795 to 2699 cal yrs BP.

The calibration distributions for the dates from the MUH2 core sequence are shown in Fig. 5.14, and these again highlight the complex multi-modal probability distributions resulting from calibration, including split distribution (MUH2-1 65-66cm; Fig. 5.14a) and very wide distribution (MUH2-4 271-272 cm; Fig. 5.14g). The latter date (MUH2-4 271-272 cm) has a conventional $^{14}$C date of $2463 \pm 24$ $^{14}$C yrs BP, and so it encounters a flattening of the radiocarbon calibration curve between about 2500 and 2400 $^{14}$C years BP (Reimer et al., 2009). The 2$\sigma$ calibrated age ranges for the MUH2 dates were plotted against depth (Fig. 5.15) to show the overall distribution of dates with depth for the core sequence. The dates for the MUH2 cores appear to be broadly in sequence, with no apparent age reversals, although some of the calibrated ranges are relatively large.
Fig. 5.14: 2σ calibration of the seven radiocarbon dates from the MUH2 core sequence, showing the resultant probability distributions. All dates were calibrated against IntCal09 using OxCal v. 4.1.7.
Based on the radiocarbon dating results, both the MUH1 and MUH2 core sequences extend back to between ca. 3000 and 2500 cal yrs BP. However, there are a few uncertainties associated with the sediment chronologies for the two core sequences, based both on the dating information itself and on observations during sediment core extrusion:

i. the two topmost cores (MUH1-1 and MUH2-1) had the highest retention of water in the coring barrel, and were also the only two cores to experience core expansion (rather than core shortening), which may indicate slumping or mixing of sediment in these cores. This was considered a more severe problem for MUH1-1 (which expanded by 27 cm) than for MUH2-1 (which expanded by 6 cm);

ii. the multi-modal probability distributions of the calibrated dates make it difficult to define a narrow age range for a particular sediment depth;

iii. the measured depths of the overlapping sediment cores during extraction may not accurately reflect their relative sediment depths within each sequence. A 3-cm offset appears to exist between the measured depths of cores MUH2-3 and MUH2-4, based on the degree of similarity between the sediment stratigraphies (Fig. 5.7); however, the offset could be even larger for the cores near the top of the sequence,
for which the sediment was more unconsolidated and had a higher water content;

iv. the relationship between the basal date of core MUH4 (1904 AD ± 20 at 35.5 cm,

as determined by the $^{210}$Pb and $^{137}$Cs chronology) and the topmost dates of the

radiocarbon chronologies for MUH1 and MUH2 is unknown.

The MUH2 cores apparently constitute a more complete sequence for the upper part of the

record, as they have longer sections of overlapping sediment, and the order of radiocarbon

dates is more consistent than for MUH1 (cf. Fig. 5.13 and Fig. 5.15). Of the five

radiocarbon dates from MUH1, one shows a slight age reversal (MUH1-3 181-182 cm),

and one may be an anomalously young age from a core with suspected sediment slumping

or mixing (MUH1-1 92-93 cm). Bayesian age models for both sequences (MUH1 and

MUH2) are described in Section 5.2.5. Physical proxy measurements of the core sediments

are first presented, however, in order to improve correlation of overlapping cores.

5.2.4 Physical proxy measurements

Results of the physical proxy measurements (%LOI, % dry weight, wet density and

magnetic susceptibility) for the MUH1 cores are plotted in Fig. 5.16. The results for core

MUH4 are also plotted, although magnetic susceptibility was not measured for this core.

The %LOI values for the MUH1 cores range between 23% and 75%, with one clear drop

in values at 137 cm in MUH1-2, and a downward trend at the bottom of MUH1-1. % dry

weight shows a generally decreasing trend towards the top of the sequence, and a marked

peak at 137 cm (MUH1-2). Wet density remains relatively stable throughout the sequence,

fluctuating between about 1 and 1.25 g cm$^{-3}$, with one marked peak of 1.5 g cm$^{-3}$ at 227 cm

(MUH1-3). Magnetic susceptibility, which was measured at a lower resolution than the

other measurements, shows background levels close to zero, and two peaks at 227 cm

(MUH 1-3) and 137 cm (MUH1-2). These two peaks both correspond to changes in the

other three proxy measurements at these depths, and indicate an increase in minerogenic

content, which was also observed in the visual descriptions of the sediment cores (cf. Fig.

5.7). However, there are no clear correlations of values of any of the proxy measurements

between similar depths in the three overlapping MUH1 cores. This lack of correspondence

of features between the overlapping cores may be problematic, as it may indicate that the
MUH1 cores cannot be combined to form a continuous, overlapping sequence. %LOI values for MUH4 increase to 85% at 31 cm from a bottommost (35 cm) value of 45%, before gradually decreasing to a surface sample value of 58%. % dry weight is generally low for this core, between 3% and 10%, reflecting the high water content observed during extrusion.

The results of the physical proxy measurements for the MUH2 cores are shown in Fig. 5.17. The data from MUH4 are also plotted. For MUH2, the %LOI values mostly range between about 40% and 60%, indicating organic-rich sediment. However, there are marked drops in %LOI values to below 20% at four depths: MUH2-4 256-257 cm, MUH2-3 253-254 cm, MUH2-2 130-131 cm, and MUH2-1 80-90 cm. % dry weight shows a gradually decreasing trend upwards, from 20% to 11%, confirming the observation that water content

![Fig. 5.16: Physical proxy measurements for the MUH1 and MUH4 cores. Calibrated radiocarbon age ranges associated with the MUH1 cores are shown.](image-url)
decreased with increasing depth in this core sequence. Four major peaks in % dry weight occur at the same depths as the drops in %LOI referred to above. Wet density also shows changes at these four depths, and otherwise stable values, although the peak at MUH2-1 80-90 cm is not particularly marked, and there is a drop, rather than a rise, in wet density values at MUH2-4 256-257 cm. Magnetic susceptibility, which was measured at a higher resolution than the other measurements, has background values close to zero, but shows significant peaks at the same four depths as the changes in the other proxy values. The highest values occur at MUH2-4 256.5 cm and MUH2-3 253 cm. A small peak also occurs at MUH2-1 108 cm; a depth for which no changes in the other proxy values are registered.

Fig. 5.17: Physical proxy measurements for the MUH2 and MUH4 cores. Calibrated radiocarbon age ranges associated with the MUH2 cores are also given, and are colour-coded according to the key shown.
The sediment in both the MUH1 and MUH2 core sequences thus consists primarily of organic-rich, homogenous lake mud, for which the water content generally declines with depth. A few sharp fluctuations in sediment content occur, consisting of abrupt increases in minerogenic content, and usually also associated with a change in colour and/or grain size (cf. Fig. 5.7). However, these fluctuations do not always occur at the same depth in overlapping cores. The lack of correspondence of features within the overlapping cores of the MUH1 core sequence, and the potential problems with the radiocarbon dating chronology, mean that building a single stacked sequence for the MUH1 cores was problematic.

Similarities in physical proxy measurements and sediment stratigraphy for cores MUH2-3 and MUH2-4 indicate that the two cores represent overlapping sections of the same sediment sequence, but with an offset of 3 cm (i.e. core MUH2-3 253 cm is equivalent to core MUH2-4 256 cm). The overlap for cores MUH2-1 and MUH2-2, and indeed for cores MUH2-2 and MUH2-3, is not so straightforward. One possibility is that the observed features in the physical proxy values (and occurrence of coarse-grained sediment) centred on MUH2-1 85 cm and MUH2-2 130 cm actually represent the same feature in the sedimentary sequence. If these were two discrete and separate fluctuations in the sediment content, then the lower fluctuation, at MUH2-2 130 cm, would be expected to also occur in core MUH2-1, which extends to a depth of 150 cm. The top of core MUH2-2 is at a measured sediment depth of 125 cm, and the short-lived change in sediment stratigraphy observed from 130-131 cm may reflect compression of the feature observed in core MUH2-1 from 80-90 cm. Conversely, core MUH2-1 may have been subject to mixing and diffusion of the coarse-grained sedimentary horizon. The physical proxy evidence thus suggests that cores MUH2-1 and MUH2-2 may be correlated based on this potentially corresponding sedimentary feature. The implications of this for chronological control are discussed in the next section.

5.2.5 Age model production and core correlation techniques

A P-sequence age model for the MUH1 core sequence was produced. No changes were made to the MUH1 measured sediment depths upon extraction. For the MUH2 core sequence, two P-sequence age models were developed (see Table 5.7). The first age model
(MUH2 Model 1) was based upon sediment depths that corresponded to the measured depths for sediment cores on extraction, with the exception that 3 cm was added to all MUH2-3 depths in order to correlate this core with core MUH2-4. The second age model (MUH2 Model 2) additionally involved bringing the sedimentary feature at 130 cm in core MUH2-2 in line with that occurring at 85 cm in core MUH2-1. Cores MUH2-2 and MUH2-3 were therefore moved up by 45 cm, and core MUH2-4 was moved up by 48 cm (Table 5.7). For all age models, the top and bottom dates of the MUH4 chronology (based on \(^{210}\)Pb and \(^{137}\)Cs dating) were also included, consisting of a surface sediment boundary dated from -53 to -58 cal yrs BP (2003 to 2008 AD), and another boundary placed at 35.5 cm (the basal \(^{210}\)Pb date) with a normally distributed date of 44 ± 20 cal yrs BP (i.e. 1906 AD ± 20). A third, undated boundary was placed at the base of each core sequence.

The results of the initial P-sequence model for the MUH1 core sequence, with \(k = 0.1\), are shown in Table 5.8. As highlighted in the table, the overall agreement index (A\(_{\text{overall}}\)) for the model is 50%, which is below the 60% threshold suggested by Bronk Ramsey (2008). All dates have a high agreement index, with the notable exception of MUH1-3 181-182 cm (the date which shows an age reversal). The low agreement index for this date (17%) is responsible for lowering A\(_{\text{overall}}\) to a value <60%. Fig. 5.18 shows the age–depth curve produced for this model, and comparisons of the modelled and unmodelled calibrated distributions of the dates are also shown.

**Table 5.7:** Sediment depths used in Models 1 and 2 for the MUH2 core sequence. Sediment depths that differ from those measured during sediment extraction are highlighted in red.

<table>
<thead>
<tr>
<th>Core</th>
<th>Measured depth of sediment extraction</th>
<th>MUH2 Model 1 sediment depth</th>
<th>MUH Model 2 sediment depth</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>top (cm)</td>
<td>base (cm)</td>
<td>top (cm)</td>
</tr>
<tr>
<td>MUH2-1</td>
<td>50</td>
<td>150</td>
<td>50</td>
</tr>
<tr>
<td>MUH2-2</td>
<td>125</td>
<td>225</td>
<td>125</td>
</tr>
<tr>
<td>MUH2-3</td>
<td>168</td>
<td>268</td>
<td>171</td>
</tr>
<tr>
<td>MUH2-4</td>
<td>235</td>
<td>292</td>
<td>235</td>
</tr>
</tbody>
</table>

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Table 5.8: OxCal model output for a P-sequence model for the MUH1 core sequence. Overall agreement index ($A_{\text{overall}}$) is 50% ($k = 0.1$). The agreement index for each modelled date is also shown ($A$).

<table>
<thead>
<tr>
<th>Name</th>
<th>Show all structure</th>
<th>Unmodelled (BP)</th>
<th>Modelled (BP)</th>
<th>Indices</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td></td>
<td>from</td>
<td>to</td>
<td>%</td>
</tr>
<tr>
<td>U(2003,2008)</td>
<td></td>
<td>-53</td>
<td>-58</td>
<td>95.4</td>
</tr>
<tr>
<td>Boundary top</td>
<td></td>
<td>-53</td>
<td>-58</td>
<td>95.4</td>
</tr>
<tr>
<td>C_Date base 210Pb</td>
<td></td>
<td>85</td>
<td>5</td>
<td>95.4</td>
</tr>
<tr>
<td>Boundary base 210Pb</td>
<td></td>
<td>85</td>
<td>5</td>
<td>95.4</td>
</tr>
<tr>
<td>R_Date MUH1-1-92cm</td>
<td></td>
<td>467</td>
<td>307</td>
<td>95.4</td>
</tr>
<tr>
<td>R_Date MUH1-2-142cm</td>
<td></td>
<td>1928</td>
<td>1727</td>
<td>95.4</td>
</tr>
<tr>
<td>R_Date MUH1-3-181cm</td>
<td></td>
<td>2705</td>
<td>2358</td>
<td>95.4</td>
</tr>
<tr>
<td>R_Date MUH1-3-210cm</td>
<td></td>
<td>2348</td>
<td>2160</td>
<td>95.4</td>
</tr>
<tr>
<td>R_Date MUH1-3-253cm</td>
<td></td>
<td>2843</td>
<td>2541</td>
<td>95.4</td>
</tr>
<tr>
<td>Boundary</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
</tbody>
</table>

The failure of the model to achieve an $A_{\text{overall}}$ of >60%, even with $k$ as low as 0.1, is a further indication of the problems associated with the apparent age reversal obtained for the date from MUH1-3 181-182 cm. Also, as the model has a very low $k$, most of the modelled distributions of dates are not much different from their unmodelled, calibrated distributions, so the model does not provide narrower age estimates (with the exception of MUH1-3 210-211 cm – see Fig. 5.18d). Removing the date with the lowest agreement index (i.e. MUH1-3 181-182 cm), which may be considered an outlier, would improve the model and allow $k$ to be increased to a value of 0.4 ($A_{\text{overall}}$ remains >60%). However, the date at MUH1-3 210-211 cm could also be considered a potential outlier, as it is younger than expected based on the MUH1-3 181-182 cm date – and if this date is instead removed, the model shows a similar improvement, also allowing $k$ to be increased to a value of 0.4 ($A_{\text{overall}}$ remains >60%). This shows that radiocarbon-based chronological control for MUH1 remains somewhat problematic.
Fig. 5.18: Age–depth model for the MUH1 core sequence dates, using a P-sequence model with $k = 0.6$. The base of the $^{210}$Pb dates for MUH4 were input as a normally-distributed age of 1904 AD ± 20 at 35.5 cm. Details of five of the modelled dates are shown as examples of how the modelled distributions compare with the unmodelled, calibrated distributions.
The results for MUH2 Model 1 are shown in Table 5.9, and plotted in Fig. 5.19. $A_{\text{overall}}$ for a P-sequence model with $k = 1$ is given as 66%. A further increase in $k$, to 1.1, yielded an $A_{\text{overall}}$ of 58%, so $k = 1$ is taken as the optimum $k$ value for the final MUH2 Model 1, as $A_{\text{overall}}$ is >60%. Comparisons of some of the modelled dates with their unmodelled, calibrated distributions are also shown in Fig. 5.19. For most of the dates (with the exception of MUH2-2 153-154 cm), the model provides a slightly narrower estimate of age than the unmodelled, calibrated distributions. For example, the date from MUH2-4 271-272 cm (Fig. 5.19e) has a very wide unmodelled age range of 2705 to 2364 cal yrs BP, due to its interception of the flattened area of the calibration curve at around 2500 $^{14}$C yrs BP, and the model narrows this range slightly to an estimate of between 2694 and 2460 cal yrs BP. The split distribution of the date from MUH2-1 65-66 cm, which is a marked feature of the unmodelled, calibrated distribution (Fig. 5.19a, and cf. Fig. 5.14a), is improved slightly by the model, but the distribution is still split between 621-606 cal yrs BP and 562-517 cal yrs BP, albeit with a much lower probability (5%) assigned to the earlier of these two age ranges. The only date with an agreement index of <60% is MUH2-2 153-154 cm (44%), as the model estimates its age as older than the calibrated distribution (Fig. 5.19c).

Table 5.9: OxCal output for MUH2 Model 1, a P-sequence model for the MUH2 core sequence. Overall agreement index ($A_{\text{overall}}$) is 66% ($k = 1$). The agreement index for each modelled date is also shown (A).
MUH2 Model 1 P-sequence ($k = 1$, $A_{\text{over}} = 66\%$)


Fig. 5.19: Model 1 for the MUH2 core sequence dates, using a P-sequence model with $k = 1$. The base of the $^{210}$Pb dates for MUH4 were input as a normally-distributed age of 1904 AD ± 20 at 35.5cm. Details of five of the modelled dates are shown as examples of how the modelled distributions compare with the unmodelled, calibrated distributions.
The results for MUH2 Model 2 (which differ from Model 1 in that cores MUH2-2, MUH2-3 and MUH2-4 were all moved upwards in order to account for the matching of sediment characteristics – see Table 5.7) are shown in Table 5.10. A P-sequence model with \( k = 1 \) yielded an \( A_{\text{overall}} \) of 63.5%. Increasing \( k \) to 1.1 yielded an \( A_{\text{overall}} \) of 59.4%, so \( k = 1 \) is taken as the optimum \( k \) value for the final MUH2 Model 2, as \( A_{\text{overall}} > 60\% \). The model is plotted in Fig. 5.20, along with comparisons of some of the modelled dates with their unmodelled, calibrated distributions. As with MUH2 Model 1, Model 2 provides a narrower age range estimate than the unmodelled, calibrated dates, and also provides a slightly narrower estimate for each date than that obtained by Model 1. The split distribution of the date at MUH2-1 65-66 cm is not, however, much improved by Model 2 (Fig. 5.20a): the modelled estimate is almost identical to that given by Model 1. No other marked split distributions are apparent in Model 2: for example, MUH2-2 153-154 cm shows a split distribution in Model 1, but in Model 2 (in which it plots at a depth of 108-109 cm) it shows an improved distribution of 1373-1302 cal yrs BP, albeit with a poor agreement (\( A = 45.3\% \)) between the unmodelled and modelled estimates (Fig. 5.20b). Model 2 places the base of the sequence (244 cm; 3089-2730 cal yrs BP) at a slightly older date than the basal estimate of Model 1 (292 cm; 2949-2622 cal yrs BP).

**Table 5.10:** OxCal output for MUH2 Model 2, a P-sequence model for the MUH2 core sequence. Overall agreement index is 63.5% (\( k=1 \)). The agreement index for each modelled date is also shown (A).
Fig. 5.20: Model 2 for the MUH2 core sequence dates, using a P-sequence model with \( k = 1 \). The base of the \(^{20}\text{Pb}\) dates for MUH4 were input as a normally-distributed age of 1904 AD ± 20 at 35.5cm. Details of five of the modelled dates are shown as examples of how the modelled distributions compare with the unmodelled, calibrated distributions.
A comparison of MUH2 Models 1 and 2 shows that based on the model outputs, the $k$ values are the same for both models, and either model could potentially be used to characterise the age/depth relationship of the sediments. However, utilising the physical proxy data in addition to the chronological evidence is a more thorough way to establish the core correlation relationship (Aaby and Berglund, 1986; Dearing, 1986; Zolitschka et al., 2001). Model 2 utilises the evidence provided by the physical proxy information, as described in Section 5.2.4, and is well-supported by the chronological evidence, as shown by the model’s viability. Model 2 is thus selected as the best estimate of core correlations for the MUH2 cores. Fig. 5.21 shows the physical proxy measurements for the MUH2 cores plotted according to Model 2 depth, with a secondary y-axis showing the Model 2 chronology. This highlights that overall, the sediment sequence from the crater lake on

Fig. 5.21: Physical proxies for the MUH4 and MUH2 core sequence plotted according to Model 2 depth, with a secondary y-axis in cal yrs BP.
Mt. Muhavura shows two marked peaks in physical proxy measurements, centred on 208 cm (ca. 2420 cal yrs BP) and 85 cm (ca. 880 cal yrs BP).

In summary, due to problems with the between-core correlations of the MUH1 sequence, and an age reversal in the radiocarbon dating, this core sequence was not used in subsequent laboratory analyses of biological and chemical proxy indicators. The sedimentary record from the crater lake on Mt. Muhavura is best represented by a combination of the MUH4 and MUH2 cores, correlated according to a combination of stratigraphical and chronological evidence (Model 2). The main physical proxies for the combined MUH4 and MUH2 sequences are summarised in Fig. 5.21. All biological and chemical proxy measurements presented in next chapter are based on the analysis of sediments from the MUH4 and MUH2 sequences, and are plotted as a single stacked sequence correlated according to Model 2.
Chapter 6 Results II: Biological and chemical proxies

This chapter presents the results of the biological and chemical proxy measurements carried out on the sediments from the crater sites on Mt Gahinga and Mt. Muhavura. Results of pollen, macrocharcoal and tephra analyses for the sediments from the crater swamp on Mt. Gahinga are presented first, followed by those for the crater lake on Mt. Muhavura. Results for the latter site also include microcharcoal, diatom and detailed geochemical data.

6.1 Mt. Gahinga crater swamp

6.1.1 Pollen analysis

A total of 73 pollen types were identified in 27 samples from the Mt. Gahinga core sequence analysed for pollen content. Pollen concentrations were generally low, and up to five slides were counted per sample in order to reach a total of 500 grains. Preservation of pollen grains was generally good. A small number of unknown and degraded grains was recorded, which amounted to <5% of the total grains counted in each sample. Fig. 6.1 shows photographs of the main taxa found in the core sequence.

Fig. 6.2 shows the percentage abundances of all pollen taxa that reach a minimum of 1% of the total pollen sum (a total of 41 taxa) in the analysed samples. Pollen percentage values in Fig. 6.2 were calculated based on the total number of grains counted, excluding unknown and degraded pollen grains. Total pollen concentration and pollen accumulation rate for each sample are also shown, expressed as pollen grains per gram dry weight of sediment, and pollen grains per cm$^3$ per year, respectively. Pollen taxa are divided into categories according to whether they are mostly associated with montane forest tree species, or scrub/herbaceous species. Categories of other arboreal species (i.e. not occurring primarily in montane forest), other spores, and aquatic/local taxa are shown on the diagram. A summary of the cumulative category percentages is also shown, highlighting that local/aquatic taxa fluctuate between ca. 5% and 40% of the total pollen sum.
Fig. 6.1: Photographs of the main pollen taxa found in the Mt. Gahinga swamp core sequence. 

Fig 6.2: Pollen diagram for the Mt. Gahinga core sequence, with all taxa included in the pollen sum. An exaggeration factor of x5 (dotted pattern) is shown for less abundant taxa. Summary graph of the total percentage of each group, and total pollen concentration and accumulation rate are also shown.
The inclusion of local taxa in the calculation of pollen percentages can lead to the over-representation of pollen from plants growing close to the coring site and to large inter-sample variations in abundance, particularly for swamp sediments (Hamilton, 1982). Pollen from plants that are likely to have been growing directly on the swamp surface may be over-represented, and include here Cyperaceae, *Pteridium, Sphagnum* and *Typha*. Pollen percentages and concentrations were therefore also calculated based on a total that excluded the aquatic/local taxa category (Fig. 6.3). Taxa reaching a minimum abundance of 1% in at least one sample (44 taxa) are shown on the diagram. A comparison of Fig. 6.2 and Fig. 6.3 shows that there appear to be no marked differences in the patterns of percentage variations in the non-aquatic/local taxa, regardless of whether the aquatic/local taxa group is included in the pollen sum. However, in order to circumvent over-representation effects in the % calculations (for example, pollen from Cyperaceae varies from 8% to 38%), the pollen sum that excludes pollen from taxa that are likely to have been growing at or close to the coring site (Fig. 6.3) can perhaps be regarded as the more accurate representation of variation in pollen from non-local (i.e. terrestrial) sources.

The results of stratigraphically constrained cluster analysis (CONISS) are plotted as a dendrogram on Fig. 6.3. CONISS was carried out based on the 44 taxa that reached an abundance of at least 1% after aquatic/local taxa had been excluded. The broken-stick model applied to the CONISS output identified two significant breaks which accounted for a greater proportion of variance than for a random dataset (Table 6.1; Fig. 6.4). These two breaks can be accorded the status of pollen zone boundaries. The first of these boundaries occurs between the samples taken from 290 and 270 cm, which was assigned a depth of 280 cm (ca. 990 cal yrs BP). The second break occurs between 650 and 590 cm (i.e. 620 cm, ca. 4500 cal yrs BP). This break is only just about statistically significant (Table 6.1; Fig. 6.4). The next level of zonation (three breaks/four zones, which according to the CONISS would occur at a depth of 96 cm) is not significant; i.e. the proportion of variance accounted for by this level of zonation in the cluster analysis is less than that expected for a random dataset, according to the broken-stick model. Three zones are therefore defined for this pollen sequence, termed Gahinga Pollen Zones (GPZs), as delimited by the first two breaks in the CONISS output (Fig. 6.3 and Fig. 6.4).
Fig. 6.3: Pollen diagram for the Mt. Gahinga core sequence, with aquatic/local taxa excluded from the total pollen sum. An exaggeration factor of x5 (dotted pattern) is shown for less abundant taxa. Summary graph of the total percentage of each group, and total pollen concentration and accumulation rate are plotted (based on the total, excluding aquatic/local taxa). CONISS dendrogram and zonation of pollen data are also shown. GPZ = Gahinga Pollen Zone.
Table 6.1: Variance accounted for by the 1\textsuperscript{st}, 2\textsuperscript{nd} and 3\textsuperscript{rd} breaks of a broken-stick model of a randomised dataset of 27 samples compared with values for the CONISS analysis of the Mt. Gahinga pollen data. The first two breaks (i.e. three zones) in the pollen data are significant.

<table>
<thead>
<tr>
<th>Number of breaks</th>
<th>Random dataset: Proportion of variance</th>
<th>CONISS: Increase in dispersion/Total dispersion</th>
<th>Depth of break</th>
</tr>
</thead>
<tbody>
<tr>
<td>1</td>
<td>0.144128</td>
<td>0.24584</td>
<td>Significant 280 cm</td>
</tr>
<tr>
<td>2</td>
<td>0.107091</td>
<td>0.107184</td>
<td>Significant 620 cm</td>
</tr>
<tr>
<td>3</td>
<td>0.088572</td>
<td>0.074917</td>
<td>Not significant (96 cm)</td>
</tr>
</tbody>
</table>

Fig. 6.4: Variance accounted for by the nth break of a broken-stick model of 27 samples compared with values for the CONISS analysis of the Mt. Gahinga pollen data. The first two breaks in the pollen data are significant (occurring at depths of 280 cm and 620 cm).

A percentage pollen diagram based on the long-distance and high-altitude pollen sums is shown in Fig. 6.5. 53 taxa and 18 taxa were classified as long-distance and high-altitude pollen types, respectively. The 43 most abundant taxa (30 long-distance and 13 high-altitude) are plotted on Fig. 6.5. The total long-distance and high-altitude pollen sums are also expressed as a cumulative percentage of the total pollen sum, and summarised in Fig. 6.5. CONISS analysis and application of the broken-stick model based only on the long-distance percentage values yielded zonations and breaks at the same depth as percentage values based on the total (minus aquatics/local taxa) sum, as summarised above in Table 6.1 and Fig. 6.4. The zonations at 620 cm (ca. 4500 cal yrs BP) and 280 cm (ca. 990 cal yrs BP) were thus also applied to the long-distance/high-altitude percentage data in Fig. 6.5.
Fig. 6.5: Long-distance and high-altitude pollen components for Mt. Gahinga core sequence. Percentage values for each pollen type are calculated based on the long-distance sum and the high-altitude sum for long-distance and high-altitude pollen types, respectively. Long-distance pollen comprises pollen types which are not associated with plants growing within high-altitude (Ericaceous or Afroalpine) vegetation. High-altitude pollen comprises pollen types which can be produced by Ericaceous or Afroalpine vegetation.
GPZ1 ranges from the basal sample from the core at 950 cm to 620 cm (ca. 7725 to 4500 cal yrs BP). The most abundant montane forest tree taxa (all of which are in the long-distance category) represented in this zone are Celtis and Olea, with Podocarpus and Rapanea also present throughout (Fig. 6.3). Other montane forest tree taxa present include Maesa, Nuxia, Polyscias and Trema. Long-distance pollen of scrub/herbaceous taxa in GPZ1 include Acalypha and Urticaceae, the latter of which maintains its highest levels (up to 15% of the total pollen sum, and up to 20% of the long-distance sum) throughout this zone. Ericaceae pollen achieves its highest relative abundance in GPZ1, reaching up to 30% of the total pollen sum (Fig. 6.3) and up to 50% of the high-altitude pollen sum (Fig. 6.5), although levels of Ericaceae pollen drop abruptly in the zone’s topmost sample (650 cm; ca. 4850 cal yrs BP). Poaceae pollen is an important component of GPZ1, remaining at ca. 10-15% of the total sum, and 20-25% of the high-altitude sum. Other high-altitude taxa in GPZ1 include Apiaceae, Asteraceae, Dendrosenecio, Hypericum and Lobelia. Pollen concentration is very low at the base of this zone, and rises to a maximum value at 774 cm (ca. 6240 cal yrs BP), corresponding with a peak in pollen accumulation rate.

GPZ2 ranges from a depth of 620 cm to 280 cm (ca. 4500 to 990 cal yrs BP). A major feature of this zone involves a significant rise in Podocarpus levels, which begin to increase initially at ca. 560 cm (ca. 3770 cal yrs BP), and then more sharply from 410 cm (ca. 2040 cal yrs BP) to the topmost sample in the zone, at 290 cm (ca. 1070 cal yrs BP), where a maximum abundance of ca. 30% of the total pollen sum and 45% of the long-distance sum is reached. Celtis and Olea remain important constituents of long-distance and montane forest tree pollen, although both begin to decrease in the upper part of the zone. Hagenia, Maesa, Phyllanthus, Polyscias, Prunus and Trema also occur at low abundances. Other notable long-distance taxa in this zone include Urticaceae pollen, which becomes much less abundant compared with GPZ1, and Acalypha, which remains at a fairly constant level throughout this part of the sequence. Ericaceae pollen is much less abundant in this zone than in GPZ1, generally remaining below 20% of the high-altitude sum (Fig. 6.5). A short-lived increase in Alchemilla, Apiaceae and Asteraceae occurs in the lowermost sample of GPZ2, and is followed by an increase in Poaceae, which remains relatively high throughout the rest of this zone, reaching up to 30% of the scrub/herbaceous
sum (Fig. 6.3) and 50% of the high-altitude sum (Fig. 6.5). Pollen concentration and accumulation rate remain relatively consistent, although there are slight peaks in both values at 410 cm (ca. 2070 cal yrs BP). The relative proportion of montane forest tree pollen and scrub/herbaceous pollen remains similar throughout GPZ1 and GPZ2. The relative proportion of long-distance and high-altitude taxa also remains similar, although with a slight increase in long-distance taxa towards the top of GPZ2.

GPZ3 extends from a depth of 280 cm to the uppermost pollen sample analysed at 62 cm (i.e. from ca. 990 to 30 cal yrs BP). The onset of this zone is characterised by an abrupt decline in *Podocarpus*, the most abundant montane forest tree and long-distance taxon throughout GPZ2. *Celtis* and *Olea* also decline in abundance in this zone. *Macaranga* occurs in the early half of GPZ3 before disappearing again, while *Polyscias*, *Prunus* and *Rapanea* continue to occur at low levels throughout GPZ3. *Syzygium* increases in the topmost sample, but is virtually absent prior to this. *Hagania* and *Myrica* both increase, from previously very low levels, to become important components of pollen from montane forest trees and of the long-distance pollen sum. The increase in *Myrica* pollen in GPZ3 is particularly marked, as it occurs only rarely in GPZ2. *Hagania* is present at low levels through GPZ1 and GPZ2, but increases markedly in GPZ3. *Dodonaea* also increases at the beginning of GPZ3, to a maximum at 230 cm (ca. 770 cal yrs BP). *Rumex* also increases in GPZ3, particularly in the upper half of the zone, and Amaranthaceae/Chenopodiaceae pollen shows a similar pattern. Urticaceae pollen declines to <3% of the total sum at the beginning of the zone, but increases again slightly towards the top, and *Acalypha* increases in the top half of the zone. As a percentage of the total pollen sum, Poaceae pollen shows a marked relative increase (Fig. 6.3), but this is mostly an artefact of the major relative decline in *Podocarpus*; as a percentage of the high-altitude sum, Poaceae levels remain much more constant (Fig. 6.5). Pollen concentration is overall slightly lower than in GPZ2, except for a peak in values at 202 cm (ca. 650 cal yrs BP). Pollen accumulation rate increases dramatically in the uppermost sample. The relative percentage of pollen from montane forest trees (primarily due to the fall in *Podocarpus*) decreases dramatically in GPZ3, with a relative increase in pollen from scrub/herbaceous sources; the total long-distance pollen percentage shows a similar relative decrease in this zone.

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Fig. 6.6: Pollen concentration for each pollen type (expressed as grains per gram dry weight of sediment), arranged according to long-distance/high-altitude grouping. Total pollen concentration and accumulation rate are also shown, indicating the relative cumulative proportion of long-distance and high-altitude taxa. PAR = pollen accumulation rate; GPZ = Gahinga Pollen Zone.
Concentration values for each of the most abundant pollen taxa are plotted in Fig. 6.6, and arranged according to the long-distance/high-altitude categories. The absolute abundances of high-altitude pollen types are generally quite high, which is consistent with a relatively local provenance for these pollen taxa. Cyperaceae, Ericaceae and Poaceae achieve higher concentrations than all other taxa except *Podocarpus*—a long-distance type with a particularly high relative export ability. The increase in *Podocarpus* is shown to be most marked from ca. 410 cm (ca. 2040 cal yrs BP) onwards. Throughout the sequence, the total long-distance and total high-altitude pollen concentration and accumulation rate tend to fluctuate concurrently, highlighting that any increase or decrease in values reflects a change in the total from both categories. Pollen accumulation rate shows a particularly marked increase in the topmost sample of the sequence.

Ordination of the Mt. Gahinga pollen data was carried out on all taxa that reached a minimum abundance of 1% in at least one sample (in order to minimise the effect of rare species). This calculation was based on the total pollen sum (including aquatic/local taxa). DCA of the pollen data gave a first axis gradient length of 1.285, which is <2 and therefore indicates that linear ordination methods are most appropriate for this dataset (ter Braak, 1995). The linear ordination method of PCA was carried out, with Aitchison’s log-ratio method applied to the data (Aitchison, 1990). The first axis of the PCA (i.e. the first latent variable identified) accounts for 27.2% of the total variance in species data (Table 6.2). The first two PCA axes together account for less than half of the total variation, which means that there is a large amount of variation in the data that is not explained by the ordination analysis (ter Braak, 1995). Nonetheless, PCA can still provide a useful way of visualising the data without stratigraphical constraints and examining the relationships and correlations between samples and species.

Table 6.2: Eigenvalues and % variance of each of the four axes of a PCA of the Mt. Gahinga pollen data. Aitchison’s log-ratio method was applied. Species that did not achieve at least 1% in any sample were excluded from the analysis, in order to minimise the effect of rare species.

<table>
<thead>
<tr>
<th>Axis</th>
<th>1</th>
<th>2</th>
<th>3</th>
<th>4</th>
</tr>
</thead>
<tbody>
<tr>
<td><strong>Eigenvalues</strong></td>
<td>0.272</td>
<td>0.148</td>
<td>0.093</td>
<td>0.082</td>
</tr>
<tr>
<td><strong>Cumulative % variance of species data</strong></td>
<td>27.2</td>
<td>42.0</td>
<td>51.3</td>
<td>59.6</td>
</tr>
</tbody>
</table>
Fig. 6.7, based on the results of PCA analysis of the pollen data from Mt. Gahinga shows the relationship between species (arrows) and samples (circles). The distance between samples on the PCA biplot is an indication of the dissimilarity of their species composition (measured by their Euclidean distance) (ter Braak and Šmilauer, 2002). Samples from the same pollen zone (as determined by the stratigraphically constrained CONISS analysis) are shown to tend to cluster together on the ordination biplot, indicating the similarity of composition of these samples, and confirming their similarity in the absence of
stratigraphical constraints. The species arrows indicate the direction of species increase: the direction and length of the arrows indicates the importance of a taxon in the composition of samples in a particular area of the ordination plot. Angles between vector arrows approximate their correlations (positive correlation when the angle is acute; negative correlation when the angle is obtuse), although it is actually the location at which an arrow tip projects at right angles onto another species arrow that indicates their correlation. Species with arrows at right angles to each other are therefore poorly correlated. From the ordination diagram, *Celtis*, Ericaceae and Urticaceae are positively correlated, and are important in the composition of GPZ1. *Podocarpus* is an important component of species composition in GPZ2, but is poorly correlated with those for GPZ1. *Myrica* is moderately well correlated with *Dodonaea*, *Hagenia* and *Rumex*, all of which are important the composition of GPZ3, and all but the latter of which are poorly correlated with *Podocarpus*.

### 6.1.2 Charcoal analysis

A total of 107 samples from the Mt. Gahinga core sequence was analysed for macrocharcoal content: 77 of these were found to contain no charcoal (sample depths are listed in Table 6.3), and 30 samples contained macrocharcoal, but at low concentrations.

<table>
<thead>
<tr>
<th>Depth (cm)</th>
<th>82</th>
<th>94</th>
<th>104</th>
<th>120</th>
<th>136</th>
<th>152</th>
<th>168</th>
<th>184</th>
<th>200</th>
<th>208</th>
<th>216</th>
<th>220</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>224</td>
<td>226</td>
<td>227</td>
<td>228</td>
<td>229</td>
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<td>235</td>
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<td>238</td>
<td>240</td>
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<td>264</td>
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<td>376</td>
<td>392</td>
<td>408</td>
<td>424</td>
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<td>642</td>
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<td>712</td>
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<td>856</td>
<td>872</td>
<td>824</td>
<td><strong>832</strong></td>
<td>836</td>
<td>838</td>
<td>895</td>
<td><strong>904</strong></td>
<td><strong>968</strong></td>
</tr>
</tbody>
</table>

**Table 6.3:** Sample depths analysed for macrocharcoal content from the Mt. Gahinga core sequence by researchers at the University of New South Wales, Sydney, Australia. Underlined, bold type indicate that charcoal was present in that sample; italics indicate that no charcoal was present.
(Fig. 6.8). The highest concentration and accumulation rate occurs at 232 cm (ca. 780 cal yrs BP), and a clustering of peaks occurs around this depth (at 227 cm, 234 cm, 237 cm and 240 cm). However, several samples which were also analysed from around this depth did not contain any charcoal (including a second sample analysed from 232 cm; see Table 6.3). There are no clear reasons why the amount of macrocharcoal in the sediment is so variable at this depth. Possibly, concentrations are so small that macrocharcoal may not be detectable in all samples, or may not be evenly distributed within a 1-cm horizon of the sediment core.

Fig. 6.8: Macrocharcoal concentration and accumulation rate (CHAR) in the Mt. Gahinga core sequence. Analyses were carried out by researchers at the University of New South Wales, Sydney, Australia. Sample depths at which no charcoal was found are marked by empty circles.
Other peaks in macrocharcoal content occur at 840 cm, 680 cm, 456 cm, 264 cm, 168 cm and 94 cm. These peaks are all relatively minor. The high organic content, and the acid conditions of peat deposition, may have contributed to the low amount of charcoal in the sediments. Alternatively, fires may have been relatively infrequent on the swamp surface, a distinct possibility given the local hydrological conditions, with therefore little opportunity for large fragments of charcoal to be deposited at the coring site. See Chapter 7 for further discussion of the macrocharcoal results, and comparison with the results from the Mt. Muhavura crater lake core sequence.

6.1.3 Tephra analysis

No visible tephra horizons were apparent in the Mt. Gahinga core sequence. The results of analysis for tephra content of five samples from the Mt. Gahinga core sequence are shown in Table 6.4 and Figure 6.9. Each sample consisted of a 5-cm contiguous sample, and sample depths were chosen such that a range of sample types were included, based on the results from other analyses: i.e. a sample with high macrocharcoal content (231-236 cm); two samples with high magnetic susceptibility and low %LOI (615-620 cm and 700-705 cm); and two samples with high %LOI and low magnetic susceptibility (447-452 cm and 678-683 cm), the latter of which had been found to contain macrocharcoal.

The sample from 447-452 cm was the only sample to contain no tephra shards. Two types of tephra were identified, each with different physical properties. The first type, and the dominant type, was a brown-coloured tephra with a sugary textured appearance and either

<table>
<thead>
<tr>
<th>Sample name</th>
<th>Depth range</th>
<th>Brown tephra</th>
<th>Colourless tephra</th>
<th>Total tephra</th>
</tr>
</thead>
<tbody>
<tr>
<td>GAH3_231-236cm</td>
<td>231-236 cm</td>
<td>149</td>
<td>0</td>
<td>149</td>
</tr>
<tr>
<td>GAH5_447-452cm</td>
<td>447-452 cm</td>
<td>0</td>
<td>0</td>
<td>0</td>
</tr>
<tr>
<td>GAH7_615-620cm</td>
<td>615-620 cm</td>
<td>3</td>
<td>5</td>
<td>8</td>
</tr>
<tr>
<td>GAH7_678-683cm</td>
<td>678-683 cm</td>
<td>2</td>
<td>4</td>
<td>6</td>
</tr>
<tr>
<td>GAH8_700-705cm</td>
<td>700-705 cm</td>
<td>1000</td>
<td>0</td>
<td>1000</td>
</tr>
</tbody>
</table>
a ‘blocky’ or flatter angular morphology, containing numerous micro-inclusions. The second type was a colourless tephra, generally with a flatter sharp/angular morphology. The largest peaks in total tephra occurred in two samples (231-236 cm and 700-705 cm) that were found to only contain brown tephra. The samples from 615-620 cm and 678-683 cm contained very low amounts of tephra: just eight and six shards, respectively, were found in total. Of these few shards, most were colourless tephra.

The tephra analysis thus shows that tephra shards are present in the Mt. Gahinga core sequence in a range of types of sample, although sometimes at very low amounts. There appears to be no consistent relationship between the other proxy measurements and the type and amount of tephra content.

Fig. 6.9: Results of tephra analysis of five samples from the Mt. Gahinga core sequence. All analysis was carried out by researchers at Swansea University, Wales. Note the different x-axis scales. The sample depth at which no tephra was found is marked by empty circles.
6.1.4 Carbon isotope analysis of radiocarbon-dated samples

The nine macrofossil samples from the sediments from Mt. Gahinga analysed for radiocarbon content (Section 5.1.2) also yielded $\delta^{13}$C values, which are plotted in Fig. 6.10. $\delta^{13}$C values ranged between -20 and -30%, indicative of plant taxa using the C$_3$ photosynthetic pathway (Meyers and Teranes, 2001).

![Graph showing $\delta^{13}$C values of the radiocarbon-dated plant macrofossils from the Mt. Gahinga swamp sequence.](image)

Fig. 6.10: $\delta^{13}$C values of the radiocarbon-dated plant macrofossils from the Mt. Gahinga swamp sequence.
6.2 Mt. Muhavura crater lake

6.2.1 Pollen analysis

A total of 95 pollen types were identified in the 32 samples analysed for pollen content from the sequence of sediment cores collected from the crater lake on Mt. Muhavura. Preservation of pollen grains was generally good. Pollen concentrations were much higher than for the Mt. Gahinga swamp sediments. A small number of unknown and degraded grains were recorded, which comprised <5% of each sample.

Fig. 6.11 shows the percentage abundances of all pollen taxa that reached a minimum of 1% in at least one sample (a total of 42 taxa) in the Mt. Muhavura lake sediments. Pollen percentages were calculated based on the total number of pollen grains counted, excluding unknown and degraded grains. Total pollen concentration and pollen accumulation rate for each sample are also shown. Pollen taxa are divided into categories according to whether they are mostly associated with montane forest tree species, scrub/herbaceous species, other arboreal species, other spores, or aquatic taxa. For the Mt. Muhavura crater lake pollen data, pollen likely from aquatic taxa are much less abundant than in the sediments from the swamp on Mt. Gahinga, remaining at <9% for all but two samples (12.5 cm and 20.5 cm), where levels rise to a maximum of 12%. This group was therefore retained in the total pollen sum, as it was much less likely to disproportionately influence the data.

The results of stratigraphically constrained cluster analysis (CONISS), based on the relative abundances of 42 taxa that reached a minimum of 1% in at least one sample, are plotted as a dendrogram in Fig. 6.11. The broken-stick model applied to the CONISS output identified one significant break (Table 6.5; Fig. 6.12), which occurs at a depth of 93 cm (ca. 1000 cal yrs BP). The next level of zonation (i.e. two breaks/three zones) is not significant; i.e. the proportion of variance associated with this level of zonation in the cluster analysis is less than that expected for a random dataset, according to the broken-stick model. Based on the numerical analysis, the pollen data from the crater lake on Mt. Muhavura were therefore divided into two zones, termed Muhavura Pollen Zones (MPZs), with the boundary at 93 cm (ca. 1000 cal yrs BP).
Fig. 6.11: Pollen diagram for the sediment sequence from the crater lake on Mt. Muhavura, based on the total pollen sum. The gap in the core record between the bottom of the MUH4 core (37 cm) and the top of the MUH2 cores (50 cm) is shown. An exaggeration factor of x5 (dotted pattern) is shown for less abundant taxa. CONISS dendrogram and zonation of pollen data is also shown. MPZ = Muhavura Pollen Zone.
Table 6.5: Variance accounted for by the 1<sup>st</sup> and 2<sup>nd</sup> breaks of a broken-stick model of a randomised dataset of 32 samples, compared with values for the CONISS analysis of the pollen data from the crater lake on Mt. Muhavura. The first break (i.e. two zones) in the pollen data is significant.

<table>
<thead>
<tr>
<th>Number of breaks</th>
<th>Random dataset: Proportion of variance</th>
<th>CONISS: Increase in dispersion/Total dispersion</th>
<th>Depth of break</th>
</tr>
</thead>
<tbody>
<tr>
<td>1</td>
<td>0.126828</td>
<td>0.34136</td>
<td>Significant 93 cm</td>
</tr>
<tr>
<td>2</td>
<td>0.09558</td>
<td>0.07791</td>
<td>Not significant (189 cm)</td>
</tr>
</tbody>
</table>

Fig. 6.12: Variance accounted for by the nth break of a broken-stick model of 32 samples compared with values for the CONISS analysis of the Mt. Muhavura pollen data. The first break in the pollen data is significant (occurring at a depth of 93 cm).

Pollen percentages based on the long-distance and high-altitude pollen sums are shown in Fig. 6.13. 71 taxa and 20 taxa were classified as long-distance and high-altitude pollen types, respectively. The 45 most abundant taxa (30 long-distance and 15 high-altitude) are plotted in Fig. 6.13. The cumulative percentages of long-distance and high-altitude pollen types as a proportion of the total pollen sum are also shown. CONISS analysis and application of the broken-stick model based only on the long-distance percentage values also yielded one break at 93 cm, thus supporting the division of the Mt. Muhavura core sequence into two zones, MPZ1 and MPZ2 (cf. Table 6.5 and Fig. 6.12).
Fig. 6.13: Long-distance and high-altitude pollen components for the sediment sequence from the crater lake on Mt. Muhavura. Percentage values for each pollen type are calculated based on the long-distance sum and the high-altitude sum for long-distance and high-altitude pollen types, respectively. An exaggeration factor of x5 (dotted pattern) is shown for less abundant taxa.
MPZ1 ranges from the base of the core sequence at 244 cm up to a depth of 93 cm (ca. 2850 to 1000 cal yrs BP). Montane forest tree pollen in this zone is dominated by *Podocarpus*, which accounts for between 10% and 22% of the total pollen sum (Fig. 6.11). *Podocarpus* also dominates the long-distance pollen category, accounting for up to 40% of the long-distance total. *Celtis* and *Olea* are also important components of montane forest tree pollen and long-distance pollen in MPZ1, although *Celtis* decreases slightly in the top few samples. Other minor components of long-distance and montane forest tree pollen in this zone include *Croton, Hagenia, Maesa, Nuxia, Phyllanthus, Rapanea* and *Trema*. The most abundant scrub/herbaceous taxa that are also in the long-distance category are *Acalypha* and *Urticaceae*, which occur throughout MPZ1. Poaceae pollen dominates the high-altitude category, accounting for between 35% and 60% of the high-altitude total (and ca. 20% of the total pollen sum), with the next most common pollen type, *Cyperaceae*, reaching levels of between 10% and 22% of the high-altitude total (<10% of the total pollen sum). *Alchemilla* is also an important component of high-altitude pollen in this zone, along with minor contributions from *Artemisia*, Asteraceae (including *Dendrosenecio*), *Ericaceae* and *Hypericum*. The relative proportion of montane forest tree pollen and scrub/herbaceous pollen remains similar (ca. 40% in each category) throughout MPZ1 (Fig. 6.11). Total long-distance pollen remains relatively constant throughout this zone, accounting for ca. 60% of the pollen total (Fig. 6.13). Pollen concentration and accumulation rate are highest at the bottom of the zone, at 227 cm (ca. 2650 cal yrs BP), and remain fairly stable throughout the rest of the zone.

MPZ2 ranges from a depth of 93 cm to the surface sediment (i.e. from ca. 1000 cal yrs BP to ca. 2008 AD (the date of collection of the sediment cores)). This zone is characterised by an abrupt decline in *Podocarpus* pollen, down to values of <10% of both the total and long-distance pollen sum. *Celtis* and *Olea* also declines in MPZ2, although the decrease in *Olea* is not as abrupt as that apparent for *Celtis* and *Podocarpus*. *Hagenia* and *Myrica* both increase in MPZ2, from previously very low levels. *Nuxia, Phyllanthus, Polyscias, Prunus* and *Rapanea* continue to occur at low abundances throughout MPZ2. An abrupt increase in *Dodonaea* occurs at the beginning of this zone (85 cm; ca. 880 cal yrs BP), while *Rumex* gradually increases throughout MPZ2, reaching maximum abundance in the topmost three samples. No major changes are observed in high-altitude taxa, although there is a slight
Fig. 6.14: Pollen concentration for each pollen type (expressed as grains per gram dry weight of sediment), arranged according to long-distance/high-altitude grouping. Total pollen concentration and accumulation rate are also shown, indicating the relative cumulative proportion of long-distance and high-altitude taxa. PAR = pollen accumulation rate.
increase in _Alchemilla_ in the topmost sample. Poaceae continues to dominate the high-altitude category, accounting for ca. 50% of the high-altitude sum (and ca. 25% of the total pollen sum). There is a slight decline in the total percentage of montane forest tree pollen at the onset of MPZ2 (Fig. 6.11), due primarily to the fall in _Podocarpus_ pollen, and a similar slight decline in the relative proportion of long-distance taxa. Total pollen concentration and accumulation rate decrease at the beginning of MPZ2, and peak at 20 cm (ca. 0 cal yrs BP).

Concentration values for each pollen type (arranged by long-distance/high-altitude provenance) are shown in Fig. 6.14, along with cumulative concentration and accumulation rate of the long-distance and high-altitude pollen totals. This highlights the marked decline in pollen concentration, particularly of some long-distance taxa, at the MPZ1/MPZ2 boundary, and shows that the decline in _Podocarpus_ pollen was not only a feature of the relative pollen proportions, but also consisted of a significant drop in absolute quantities of _Podocarpus_ (and _Olea_ and _Celtis_) pollen reaching the site at this time.

Ordination of the pollen data from the crater lake on Mt. Muhavura was carried out on all taxa that reached an abundance of >1% in at least one sample (in order to minimise the effects of rare species). DCA of the pollen data gave a first axis gradient length of 0.892, which is <2 and therefore indicates that linear ordination methods are most appropriate for this dataset (ter Braak, 1995). The linear ordination method of PCA was thus carried out, with Aitchison’s log-ratio method applied to the data (Aitchison, 1990). The first axis of the PCA accounts for 38.6% of the total variance in species data (Table 6.6). The first two PCA axes together account for about half (49.8%) of the total variation in the data, thus capturing a reasonable proportion of the variation in the dataset (ter Braak, 1995), and providing a useful method to examine the relationship between samples and species.

Fig. 6.15, based on the results of PCA analysis of the pollen data from the crater lake on Mt. Muhavura, shows the relationship between species (arrows) and samples (circles). Samples from the same pollen zone (as determined by the stratigraphically constrained CONISS analysis) are shown to tend to cluster together on the ordination biplot, indicating the similarity of composition of these samples, and confirming their similarity without stratigraphical constraints. The most obvious feature of the ordination biplot is the
clustering of samples from MPZ1 on the left-hand side of the diagram, and those from MPZ2 on the right-hand side, thus confirming the similarity of species composition of samples from within the same zone. This also indicates that the composition of the two zones is characterised by different pollen taxa. *Podocarpus*, *Celtis*, *Croton*, *Trema* and, to a lesser extent, *Olea*, all show a strong positive correlation, and are all important components of species composition in MPZ1. These taxa all show a relatively strong negative correlation with the main drivers of species composition in MPZ1, including *Dodonaea*, *Hagenia*, *Myrica* and *Rumex*. Overall the ordination confirms the division of the pollen diagram into two zones, MPZ1 and MPZ2, and provides a useful way of confirming the correlations between both species and samples.

Table 6.6: Eigenvalues and % variance of each of the four axes of a PCA of the Mt. Muhavura pollen data. Aitchison’s log-ratio method was applied. Species that did not achieve at least 1% in any sample were excluded from the analysis, in order to minimise the effect of rare species.

<table>
<thead>
<tr>
<th>Axis</th>
<th>1</th>
<th>2</th>
<th>3</th>
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</thead>
<tbody>
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<td>0.064</td>
<td>0.051</td>
</tr>
<tr>
<td>Cumulative % variance of species data</td>
<td>38.6</td>
<td>49.8</td>
<td>56.3</td>
<td>61.4</td>
</tr>
</tbody>
</table>

Fig. 6.15: Biplot of the PCA results for Mt. Muhavura pollen data. Samples are represented by circles; pollen taxa are represented by arrows. Sample numbers represent sediment depth (cm). Samples are colour-coded based on the zones identified by the CONISS analysis.
6.2.2 Charcoal analysis

The results of macrocharcoal and microcharcoal analyses of samples from the Mt. Muhavura core sequence are shown in Fig. 6.16. A total of 51 samples from the core sequence were analysed for macrocharcoal content, 46 of which were found to contain macrocharcoal. Macrocharcoal was thus found to occur throughout almost the entire sequence. The highest concentrations and charcoal accumulation rates occur between 84 cm and 86 cm (ca. 880 cal yrs BP), and there is a clustering of samples between 86 cm and 94 cm which were also found to contain relatively large amounts of macrocharcoal (Fig. 6.16). Other macrocharcoal peaks occur at 140 cm and 108 cm. Samples were analysed at a 1-cm resolution between the depths of 202 and 212 cm, in order to investigate whether the major change in magnetic susceptibility and %LOI at 208 cm was also related to changes in the macrocharcoal record. Only relatively low concentrations of macrocharcoal were found in these samples, however.

![Fig. 6.16: Macrocharcoal and microcharcoal results from the Mt. Muhavura crater lake sediment cores. Macrocharcoal analysis was carried out by researchers at the University of New South Wales, Sydney, Australia. Samples containing no macrocharcoal are marked by empty circles. Note the different x-axis scales.](image-url)
Results of microcharcoal analysis based on size-classing show an overall slight increase from the base of the sequence up to a maximum concentration at 134 cm (ca. 1670 cal yrs BP), and then an overall decline in values towards the top of the sequence. The point-count microcharcoal data also show an overall rise in values from the base of the sequence (to a depth of 147 cm, ca. 1830 cal yrs BP), but concentrations sustain high levels until the top of sequence, except for a dip in values between 67 and 36 cm (Fig. 6.16). Both microcharcoal records, however, show a marked degree of fluctuation, and neither show a particularly consistent pattern. Furthermore, there is a difference of two orders of magnitude in the concentration values of quantified microcharcoal between the two methods, suggesting that they may not be directly comparable.

The macrocharcoal data for the Mt. Muhavura core sequence thus show a marked pattern of discrete peaks occurring at several sample depths, whereas the microcharcoal data show a greater degree of fluctuation, with no obvious peaks, and appears to show no correlation with the macrocharcoal results. This is not surprising, as macrocharcoal and microcharcoal in sedimentary records are reflective of different spatial scales (different sizes of source area) of fire activity (Whitlock and Larsen, 2001; Mooney and Tinner, 2011). The variability in the microcharcoal data may be due to the wide area that it is likely to represent, whereas the macrocharcoal data is likely to be indicative of fire activity in a much smaller area, closer to the sampling site. Further discussion of this, and comparison with the Mt. Gahinga macrocharcoal record, is provided in Chapter 7.

6.2.3 Tephra analysis

No visible tephra horizons were apparent in the Mt. Muhavura core sequence. The results of preliminary analysis for tephra content in twenty 5-cm samples from throughout the sequence (following Pilcher and Hall’s (1992) technique) are listed in Table 6.7, and plotted on the first graph in Fig. 6.17. Possible tephra shards were identified in most of the samples, although at very low amounts. Tephra shards were brown or colourless, and usually appeared to contain vesicular micro-inclusions.

Confirmation of this preliminary tephra identification, and a more quantitative measurement of tephra content, was obtained by analysis of thirteen 5-cm samples at
Swansea University, the results of which are shown in Table 6.8 and Fig. 6.17. Two samples of 1 cm resolution (84-85 cm and 85-86 cm) were also analysed. Sample depths were selected to include samples in which no tephra had been detected by the preliminary analysis, as well as samples on which no preliminary analysis had been carried out, which were from a relatively homogenous section of the core (in terms of the physical proxy measurements). All fifteen samples were found to contain tephra. Two types of tephra were identified, with similar morphologies to those found in the Mt. Gahinga crater swamp sediments: a brown-coloured tephra with a sugary textured appearance and either a ‘blocky’ or flatter angular morphology, containing numerous micro-inclusions; and a colourless tephra with generally a flatter sharp/angular morphology. Three of the samples (84-85 cm, 85-86 cm, and 85-90 cm) contained a high concentration of shards. In particular, the concentration of tephra at 84-85 cm and 85-86 cm is extremely high, as these were only 1 cm depth samples compared to 5 cm for the other samples. All of these samples contained a high number of the brown tephra shards. Two samples (79-84 cm and 206-211 cm) contained a moderate concentration of shards (~70), while the remaining samples all contain some shards (6-36), predominantly of the colourless type.

Table 6.7: Results of preliminary tephra analysis based on Pilcher and Hall’s (1992) technique.

<table>
<thead>
<tr>
<th>Sample name</th>
<th>Depth range (cm) *</th>
<th>Possible tephra shards</th>
</tr>
</thead>
<tbody>
<tr>
<td>MUH2-1_55-60cm</td>
<td>55-60</td>
<td>3</td>
</tr>
<tr>
<td>MUH2-1_70-75cm</td>
<td>70-75</td>
<td>1</td>
</tr>
<tr>
<td>MUH2-1_84-85cm</td>
<td>84-85</td>
<td>16</td>
</tr>
<tr>
<td>MUH2-1_130-131cm</td>
<td>85-86</td>
<td>8</td>
</tr>
<tr>
<td>MUH2-1_85-90cm</td>
<td>85-90</td>
<td>15</td>
</tr>
<tr>
<td>MUH2-2_135-140cm</td>
<td>90-95</td>
<td>1</td>
</tr>
<tr>
<td>MUH2-1_100-105cm</td>
<td>100-105</td>
<td>1</td>
</tr>
<tr>
<td>MUH2-2_150-155cm</td>
<td>105-110</td>
<td>3</td>
</tr>
<tr>
<td>MUH2-1_115-120cm</td>
<td>115-120</td>
<td>5</td>
</tr>
<tr>
<td>MUH2-2_165-170cm</td>
<td>120-125</td>
<td>3</td>
</tr>
<tr>
<td>MUH2-1_130-135cm</td>
<td>130-135</td>
<td>0</td>
</tr>
<tr>
<td>MUH2-2_180-185cm</td>
<td>135-140</td>
<td>2</td>
</tr>
</tbody>
</table>

* Depth range is calculated based on Model 2 depth.

<table>
<thead>
<tr>
<th>Sample name</th>
<th>Depth range (cm) *</th>
<th>Possible tephra shards</th>
</tr>
</thead>
<tbody>
<tr>
<td>MUH2-2_195-200cm</td>
<td>150-155</td>
<td>5</td>
</tr>
<tr>
<td>MUH2-2_210-215cm</td>
<td>165-170</td>
<td>6</td>
</tr>
<tr>
<td>MUH2-3_187-192cm</td>
<td>142-147</td>
<td>1</td>
</tr>
<tr>
<td>MUH2-3_202-207cm</td>
<td>157-162</td>
<td>6</td>
</tr>
<tr>
<td>MUH2-3_217-222cm</td>
<td>172-177</td>
<td>2</td>
</tr>
<tr>
<td>MUH2-3_232-237cm</td>
<td>187-192</td>
<td>5</td>
</tr>
<tr>
<td>MUH2-4_240-245cm</td>
<td>192-197</td>
<td>0</td>
</tr>
<tr>
<td>MUH2-3_247-252cm</td>
<td>202-207</td>
<td>3</td>
</tr>
<tr>
<td>MUH2-3_253-254cm</td>
<td>208-209</td>
<td>12</td>
</tr>
<tr>
<td>MUH2-4_256-257cm</td>
<td>208-209</td>
<td>7</td>
</tr>
<tr>
<td>MUH2-4_260-265cm</td>
<td>212-217</td>
<td>1</td>
</tr>
<tr>
<td>MUH2-4_270-275cm</td>
<td>222-227</td>
<td>2</td>
</tr>
</tbody>
</table>

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Fig. 6.17: Tephra results for the sediment cores from the crater lake on Mt. Muhavura. The main tephra peak occurs at 85-86 cm, but is spread across 84-90 cm. Samples in which no tephra was found are marked with an empty circle.

Table 6.8: Results of tephra analysis of fifteen samples from the Mt. Muhavura core sequence, carried out by researchers at Swansea University.

<table>
<thead>
<tr>
<th>Sample name</th>
<th>Depth (cm)*</th>
<th>Average depth (cm)*</th>
<th>Brown Tephra</th>
<th>Colourless Tephra</th>
<th>Total Tephra</th>
</tr>
</thead>
<tbody>
<tr>
<td>MUH2-1_55-60cm</td>
<td>55-60</td>
<td>57.5</td>
<td>3</td>
<td>18</td>
<td>21</td>
</tr>
<tr>
<td>MUH2-1_79-84cm</td>
<td>79-84</td>
<td>81.5</td>
<td>73</td>
<td>2</td>
<td>75</td>
</tr>
<tr>
<td>MUH2-1_84-85cm</td>
<td>84-85</td>
<td>84.5</td>
<td>617</td>
<td>10</td>
<td>627</td>
</tr>
<tr>
<td>MUH2-2_130-131cm</td>
<td>85-86</td>
<td>85.5</td>
<td>2000</td>
<td>0</td>
<td>2000</td>
</tr>
<tr>
<td>MUH2-1_85-90cm</td>
<td>85-90</td>
<td>87.5</td>
<td>1205</td>
<td>15</td>
<td>1220</td>
</tr>
<tr>
<td>MUH2-1_106-111cm</td>
<td>106-111</td>
<td>108.5</td>
<td>3</td>
<td>24</td>
<td>27</td>
</tr>
<tr>
<td>MUH2-2_155-160cm</td>
<td>110-115</td>
<td>112.5</td>
<td>5</td>
<td>29</td>
<td>34</td>
</tr>
<tr>
<td>MUH2-2_180-185cm</td>
<td>135-140</td>
<td>137.5</td>
<td>0</td>
<td>12</td>
<td>12</td>
</tr>
<tr>
<td>MUH2-2_200-205cm</td>
<td>155-160</td>
<td>157.5</td>
<td>1</td>
<td>27</td>
<td>28</td>
</tr>
<tr>
<td>MUH2-3_220-225cm</td>
<td>175-180</td>
<td>177.5</td>
<td>10</td>
<td>26</td>
<td>36</td>
</tr>
<tr>
<td>MUH2-3_239-244cm</td>
<td>194-199</td>
<td>196.5</td>
<td>0</td>
<td>16</td>
<td>16</td>
</tr>
<tr>
<td>MUH2-3_247-252cm</td>
<td>202-207</td>
<td>204.5</td>
<td>3</td>
<td>13</td>
<td>16</td>
</tr>
<tr>
<td>MUH2-4_254-259cm</td>
<td>206-211</td>
<td>208.5</td>
<td>42</td>
<td>21</td>
<td>68</td>
</tr>
<tr>
<td>MUH2-3_260-265cm</td>
<td>215-220</td>
<td>217.5</td>
<td>2</td>
<td>12</td>
<td>14</td>
</tr>
<tr>
<td>MUH2-4_281-286cm</td>
<td>233-238</td>
<td>235.5</td>
<td>1</td>
<td>15</td>
<td>16</td>
</tr>
</tbody>
</table>

* Based on Model 2 depth.
6.2.4 Diatom analysis

In the 32 samples from the sediment sequence from the crater lake on Mt. Muhavura analysed for diatom content, a total of 52 diatoms taxa was identified. Photographs of the main diatom species enumerated in the cores are shown in Fig. 6.18. Diatom preservation was very good throughout the core sequence; very little dissolution of diatom valves was apparent. A count of 400 valves was achieved for all samples except one, from 85 cm, which was prepared from sediment from within the minerogenic layer occurring at this depth. Diatom concentration was very low in this sample, and a count of just 266 valves was achieved (which involved a count of 1792 microspheres).

The percentage abundances of the most common diatom taxa in the sediment sequence are plotted in Fig. 6.19. Total diatom concentration (per gram dry weight of sediment) is also shown. The results of stratigraphically constrained cluster analysis (CONISS), based on the abundances of 41 taxa that reached a minimum abundance of 1% in at least one sample, are plotted as a dendrogram on Fig. 6.19. The broken-stick model applied to the CONISS output (shown in Table 6.9 and Fig. 6.20) identified four significant breaks, which accounted for a greater proportion of variance than for a random dataset, occurring at sediment depths of 237 cm, 122 cm, 93 cm and 34 cm. The diatom sequence was thus divided into five assemblage zones (Muhavura Diatom Zones; MDZs), which are shown on Fig. 6.19. Diatom concentration values for each taxon were also calculated, and are plotted in Fig. 6.21 along with the total diatom concentration, total diatom accumulation rate and Hill’s N2 diversity index.

Generally, the most abundant species throughout the core sequence is *Brachysira brebissonii* (R. Ross in Hartley), with the exception of the uppermost and lowermost samples analysed. Several small *Eunotia* species occur throughout the sequence, particularly in the upper half of the core from 122 cm (ca. 1500 cal yrs BP) onwards. *Frusulcia rhomboides* (Ehrenb.) De Toni is another important component of most samples in the core. Only one centric species was enumerated, *Aulacoseira* sp., and only ever attained very low abundances.
Fig. 6.19: Diatom percentage abundances in the core sequence from the crater lake on Mt. Muhavura. Total diatom concentration (dry weight) is also shown, along with the CONISS dendrogram and the resultant zonations (MDZ1 to MDZ5) based on the application of the broken-stick model.
Table 6.9: Variance accounted for by the 1st to 5th breaks of a broken-stick model of a randomised dataset of 32 samples, compared with values for the CONISS analysis of the diatom data from the sediment sequence from the crater lake on Mt. Muhavura. The first four breaks (i.e. five zones) in the diatom data are significant.

<table>
<thead>
<tr>
<th>Number of breaks</th>
<th>Random dataset Proportion of variance</th>
<th>CONISS Increase in dispersion/Total dispersion</th>
<th>Depth of break</th>
</tr>
</thead>
<tbody>
<tr>
<td>1</td>
<td>0.126828</td>
<td>0.278526</td>
<td>Significant 122 cm</td>
</tr>
<tr>
<td>2</td>
<td>0.095578</td>
<td>0.116424</td>
<td>Significant 34 cm</td>
</tr>
<tr>
<td>3</td>
<td>0.079953</td>
<td>0.090702</td>
<td>Significant 237 cm</td>
</tr>
<tr>
<td>4</td>
<td>0.069536</td>
<td>0.086626</td>
<td>Significant 93 cm</td>
</tr>
<tr>
<td>5</td>
<td>0.061724</td>
<td>0.047112</td>
<td>Not significant (84 cm)</td>
</tr>
</tbody>
</table>

Fig. 6.20: Variance accounted for by the nth break of a broken-stick model of 32 samples compared with values for the CONISS analysis of the diatom data from the sediment sequence from the crater lake on Mt. Muhavura. The first four breaks in the diatom data are significant (occurring at depths of 122 cm, 34 cm, 237 cm and 93 cm).
Fig. 6.21: Diatom concentration values per taxa (expressed in valves per gram dry weight of sediment) for the sediment sequence from the crater lake on Mt. Muhavura. Total diatom concentration and accumulation rate is also shown, along with Hill’s N2 diversity index.
The upper zone boundary for MDZ1 occurs at 237 cm (ca. 2750 cal yrs BP). This zone consists of only one sample (from a depth of 242 cm) which has a very different assemblage composition to the samples above it. In particular, *B. brebissonii*, which has an abundance of at least 20% in all other samples, is almost absent in this sample, reaching a relative abundance of <2% (Fig. 6.19). Diatom concentration data also show that this species is almost absent in MDZ1 (Fig. 9.21) *Pinnularia subcapitata* Greg. has a 15% abundance in this zone, which is by far its highest occurrence: this species occurs in several other samples, but at much lower levels (<2%). Two other *Pinnularia* species, *Pinnularia borealis* Ehrenb. and *Pinnularia microstauron* (Ehrenb.) Cleve are important constituents of MDZ1. *F. rhomboides* also occurs in this zone, and the most abundant *Eunotia* species is *Eunotia tenella* (Grun.) Hust. Total diatom concentration is low in this zone, and Hill's N2 diversity index is relatively high, largely owing to the lack of dominance of any one taxon.

MDZ2 ranges from 237 cm to 122 cm (from ca. 2750 to 1500 cal yrs BP). This zone is characterised by high levels of *B. brebissonii*, which has a lowest abundance of 34% in the lowermost sample in MDZ2, and an abundance of >42% thereafter, with a maximum of 55% at 192 cm (Fig. 6.19). *E. tenella* is the dominant *Eunotia* species in MDZ2, and shows an overall decline towards the top of the zone, from 17% to 5%. Several other *Eunotia* species occur at low abundances throughout MDZ2. *F. rhomboides* reaches its maximum value in the lower part of MDZ2, with an abundance of 24% at 258 cm, followed by a sharp decrease to 8% at 255 cm. *Navicula mediocris* Krasske and *Navicula subtilissima* Cleve both occur at low levels in the lower half of MDZ2. *Pinnularia biceps* Greg. and *P. microstauron* both occur in MDZ2, increasing from <4% to levels between 5% and 11% from 259 cm upwards. Total diatom concentration increases in this zone, and a particularly marked increase in concentrations of *B. brebissonii*, *P. biceps* and *P. microstauron* is apparent (Fig. 6.21). Diatom accumulation rate increases in the first half of this zone before declining slightly.

MDZ3 ranges from 122 cm to 93 cm (from ca. 1500 to 1000 cal yrs BP). According to the percentage data (Fig. 6.19), *B. brebissonii* decreases slightly in this zone, but is still the dominant species, with abundances of between 35% and 43%. Concentration data,
however, show that *B. brebissonii* actually reaches its maximum absolute abundance in this zone – largely due to the peak in total diatom concentration that occurs at 110 cm (ca. 1400 cal yrs BP). *E. tenella* is much less abundant in MDZ3 than in the previous two zones, but four other *Eunotia* species increase markedly in both relative and absolute abundances – *Eunotia bilunaris* var. *mucophila* Lange-Bertalot & Nörpel, *Eunotia paludosa* var. *paludosa* Grun. in van Heurck, *Eunotia rhomboidea* Hust. and *Eunotia subarcuatoidea* Alles, Nörpel & Lange-Bertalot. This marks a diversification of the *Eunotia* species in the core. Diatom concentrations reach a marked peak in this zone at 110 cm, and decline sharply thereafter.

MDZ4 ranges from 93 cm to 34 cm (from ca. 1000 to 50 cal yrs BP). This zone is also dominated by *B. brebissonii*, although concentration data show a marked decline in abundance relative to MPZ3. This zone contains a range of *Eunotia* species, but at significantly lower concentrations than in MDZ3. *F. rhomboides* falls to very low levels, disappearing completely from the record for most of MDZ4. *P. biceps* attains its maximum relative abundance (18%), and *Stauroforma exiguiiformis* (Lange-Bertalot) Flower, Jones & Round., *Surirella engleri* O. Müller and *Aulacoseira* sp. all appear for the first time in this zone. Diatom concentration remains low throughout MDZ4, and falls to almost zero at 85 cm (ca. 880 cal yrs BP) – a depth at which Hill’s N2 shows a marked peak, followed almost immediately by an abrupt decline at 83 cm.

MDZ5, ranging from 34 cm upwards (ca. 50 cal yrs BP onwards; i.e. the past century) consists of the uppermost four samples in the core sequence. *B. brebissonii* is the most abundant species (>20% in all samples) except for in the topmost sample (2.75 cm), in which *Gomphonema parvulum* (Kütz.) Kütz. is the most abundant species (35%). Diatom concentrations increase in this zone, with a particularly rapid increase for the top couple of samples. Fig. 6.21 shows that the increase in diatom concentration consists almost completely of *G. parvulum*, with other taxa not showing a marked peak in the topmost sample.
Table 6.10: Eigenvalues and % variance of each of the four axes of a DCA of the diatom data from the sediment sequence from the crater lake on Mt. Muhavura.

<table>
<thead>
<tr>
<th>Axis</th>
<th>1</th>
<th>2</th>
<th>3</th>
<th>4</th>
</tr>
</thead>
<tbody>
<tr>
<td>Eigenvalues</td>
<td>0.305</td>
<td>0.072</td>
<td>0.051</td>
<td>0.025</td>
</tr>
<tr>
<td>Lengths of gradient</td>
<td>2.076</td>
<td>1.425</td>
<td>1.259</td>
<td>0.942</td>
</tr>
<tr>
<td>Cumulative % variance of species data</td>
<td>29.1</td>
<td>36.0</td>
<td>40.9</td>
<td>43.4</td>
</tr>
</tbody>
</table>

Fig. 6.22: DCA biplot of the diatom data from the sediment sequence from the crater lake on Mt. Muhavura. Samples are represented by circles, and diatom species are represented by triangles. Samples have been colour-coded according to zone, as identified by the CONISS analysis (cf. Fig. 6.19 and Fig. 6.20).
Ordination of the Mt. Muhavura diatom data was carried out on all taxa that reached a minimum abundance of 1% in at least one sample (in order to minimise the effects of rare species). DCA of the diatom data gave a first axis gradient length of 2.076. This is on the boundary of whether unimodal or linear methods are appropriate (ter Braak, 1995). However, linear ordination (PCA) showed a pronounced horseshoe effect, which is a distortion of the diagram that involutes the ends of the first gradient (ter Braak, 1995). DCA was therefore used to analyse the data. The DCA output (Table 6.10) shows that the first two DCA axes account for 36% of the variance in the species data.

A biplot of the DCA output is shown in Fig. 6.22. On the ordination biplot, species symbols in close proximity correspond to species that tend to occur together, and the distance between a sample point and species points approximates the (predicted) relative occurrence of species in that sample. Distance between sample symbols also approximates the dissimilarity of their species composition (measured by their Chi-squared distance), and samples where a particular species has higher frequency tend to have their symbols clustered around that species point (ter Braak and Šmilauer, 2002). Fig. 6.22 shows that samples from the same diatom assemblage zone show a tendency to cluster together on the ordination biplot, thus confirming the similarity of composition of samples from the same zone, without the stratigraphical constraints of CONISS.

6.2.5 Geochemical analysis

The δ¹³C measurements obtained for the AMS radiocarbon-dated samples from the Mt. Muhavura core sequence (Fig 6.23) range between -26.2‰ and -17.7‰. Values of less than -21‰ are consistent with having been derived from plants following the C₃ photosynthetic pathway (Meyers and Teranes, 2001). Only one sample (at 167 cm) has a value higher than -21‰; this sample consisted of a mix of several plant fragments, and therefore may have included the incorporation of some aquatic taxa, including algae (Smith and Epstein, 1971; Meyers, 1994).

The results of geochemical analysis of organic matter from the Mt. Muhavura core sequence are shown in Fig. 6.24. Percentage total organic carbon (%TOC) ranges between 1% and 33%, although most values lie between 20% and 30%, indicating high organic
Fig. 6.23: $\delta^{13}$C values of radiocarbon-dated samples from sediment sequence from the crater lake on Mt. Muhavura.

matter content (as shown also by the %LOI results). Two particularly marked downward excursions in %TOC occur at 208 cm and at 85 cm, with a smaller drop occurring at 105 cm. Percentage total organic nitrogen (%TON) shows a similar pattern: values mostly lie between 1.3% and 2.1%, but abrupt downward peaks occur at 208 cm and 85 cm, and, to a lesser extent, at 105 cm.

C/N ratios range between 12.4 and 17.6. The highest values, generally between 15.5 and 17.6, occur from the base of the core up to 94 cm (with the exception of three samples between 208 cm and 219 cm, which have slightly lower values). From 94 cm onwards, C/N ratios are all less than 15.5, with the minimum value of 12.4 occurring at 81 cm. These values are indicative of a mixed organic matter source of both higher plant material (either $C_3$ or $C_4$) and aquatic material (algal and/or aquatic macrophytes) (Tyson, 1995).
Fig. 6.24: Organic matter geochemical data for the Mt. Muhavura core sequence.
\[ \delta^{13}C \] values range between -23.54‰ and -20.17‰. A number of marked fluctuations occur in the bottom part of the core (below about 185 cm), but from 185 cm to 125 cm, values generally remain stable at around -21‰, except for a small drop to ca. -21.5‰ in two samples at 175 cm and 165 cm. From 125 cm upwards, values fluctuate markedly again, including both the highest and lowest values encountered in the sequence. \[ \delta^{15}N \] values show a similar pattern to \[ \delta^{13}C \], with a range of between 1.5‰ and 3.93‰, and with relatively stable values occurring between 185 cm and 125 cm.

A plot of C/N ratios vs. \[ \delta^{13}C \] (Fig. 6.25) shows the composition of the Mt. Muhavura crater lake samples compared with typical sources of organic matter, as outlined in Meyers (1994). The samples are consistent with a mixed source of organic matter, with a more significant contribution from \( C_3 \) plants than from \( C_4 \) plants, and also with probable inputs of algal sources.
Chapter 7 Results: Summary and analysis

This chapter presents an illustrated synthesis of the physical, chemical and biological proxy data described in the previous two chapters. Results of sedimentary analyses from both the crater swamp on Mt. Gahinga and the crater lake on Mt. Muhavura are summarised and compared, and plotted on a common age axis (in cal yrs BP), as determined by the age–depth models developed in Chapter 5 (summarised in Fig. 5.4 and Fig. 5.20). Results for the Mt. Muhavura core sequence not available for the Mt. Gahinga sequence – namely, microcharcoal data, detailed geochemical data, and diatom data – are also plotted on an age axis and described in this chapter, in order to facilitate more direct comparisons with the other proxy measurements. The timing of sedimentary changes in the records from the two sites is compared and analysed, thus providing a context for a more detailed discussion in the next chapter.

7.1 Physical and chemical proxy measurements – synthesis and comparison

A comparison of the results for %LOI, % dry weight, wet density, magnetic susceptibility, charcoal and tephra analyses from the two crater sites is plotted in Fig. 7.1. Measurements from the two sites are plotted on x-axes with the same scale, in order to facilitate direct comparisons.

The period from ca. 8000 to 2900 cal yrs BP is represented only by swamp sediments from Mt. Gahinga. There are two excursions in %LOI, % dry weight, wet density and magnetic susceptibility in this section of the core sequence, occurring at ca. 5500 and 4500 cal yrs BP. Neither macrocharcoal nor tephra were found to occur at the larger excursion at ca. 4500 cal yrs BP (although analyses for both these components were carried out at this sediment depth). The largest amount of tephra shards found in the Mt. Gahinga sequence coincides with the smaller excursion in the other proxy measurements, at ca. 5500 cal yrs BP. This tephra consisted entirely of the brown tephra type. Unfortunately, this sample was not analysed for macrocharcoal content.
Fig. 7.1: Comparison of the results of %LOI, % dry weight, wet density, magnetic susceptibility, macrocharcoal and tephra analyses for the core sequences from both study sites. Microcharcoal data (based on the size-classing results) from the core sequence from the crater lake on Mt. Muhavura are also shown. Results are plotted on a common age axis (cal yrs BP), and each set of paired graphs is plotted on the same x-axis scale in order to facilitate direct comparisons. An exaggeration factor of x5 is applied to the magnetic susceptibility and macrocharcoal data from the core sequence from the crater swamp on Mt. Gahinga G = Mt. Gahinga crater swamp; M = Mt. Muhavura crater lake.
The period from ca. 2900 cal yrs BP to the present is represented by the sediment records from the crater sites on both Mt. Gahinga and Mt. Muhavura. The Mt. Muhavura lake sediments generally show a much greater range of values for all the proxy measurements shown in Fig. 7.1. The largest excursions in %LOI, % dry weight, wet density and magnetic susceptibility in the Mt. Muhavura lake record occur at ca. 2400 cal yrs BP. As outlined in Section 5.2.1 (and illustrated in Fig. 5.7 and Fig. 5.8), the excursions in proxies coincided with a short-lived coarsening of the sediment texture and change in sediment colour. Macrocharcoal and tephra concentrations do not show major increases at this time, although low amounts of both were found to occur. The Mt. Gahinga swamp sequence does not show changes in any of the proxy measurements at this time, which indicates that the sedimentary changes from Mt. Muhavura most likely reflect a localised phenomenon at the crater lake site.

Between ca. 900 and 700 cal yrs BP, several changes are apparent in both records. %LOI in the Mt. Gahinga swamp sequence shows a slight decrease (to values of between 78% and 90%) during this time. The largest macrocharcoal peak in the Mt. Gahinga swamp sediments occurs during this interval (centred on ca. 780 cal yrs BP), coinciding with the second-largest occurrence of (brown) tephra. However, the values of % dry weight, wet density and magnetic susceptibility for the Mt. Gahinga record do not show any changes during this time. The record from the crater lake on Mt. Muhavura shows several marked changes between ca. 900 and 700 cal yrs BP, consisting of a fall in %LOI and marked increases in % dry weight, wet density, magnetic susceptibility, macrocharcoal and tephra content. The changes in values for the former four measurements are of a smaller magnitude than those that occur earlier, at ca. 2400 cal yrs BP, but the largest peaks in both macrocharcoal and tephra occur in the later excursion, centred on ca. 880 cal yrs BP. The peak in macrocharcoal concentration in the Mt. Muhavura lake sediments is several orders of magnitude higher than that for the Mt. Gahinga swamp sediments, which probably reflects the overall higher organic content and lower variability in the Mt. Gahinga swamp sediments. The relative timing of the sedimentary changes apparent at the two sites between ca. 900 and 700 cal yrs BP will be assessed further, particularly in the context of concomitant changes in pollen data, in Section 7.4.
Geochemical data measured for the Mt. Muhavura sediments are plotted on an age axis in Fig. 7.2. The δ^{13}C measurements obtained from the radiocarbon-dated samples from both sites are also shown. The two most marked drops in both %TOC and %TON occur at ca. 2400 and 900 cal yrs BP, thus coinciding with the two excursions in the other proxy measurements plotted in Fig. 7.1. C/N ratios in the Mt. Muhavura core sequence are indicative of a mixed terrestrial and aquatic source of organic matter, although the higher C/N ratios prior to ca. 1000 cal yrs BP suggest that the contribution from vascular plants was more significant during this period. This could comprise terrestrial or aquatic macrophyte contributions. After ca. 1000 cal yrs BP, the fall in C/N ratios to values close

![Graph showing geochemical data](image)

**Fig. 7.2:** Geochemical data, plotted on an age axis (cal yrs BP). Black squares and grey circles indicate δ^{13}C measurements from the radiocarbon-dated samples from the sediment sequences on Mt. Gahinga and Mt. Muhavura, respectively; all other data are from sediment organic matter in the Mt. Muhavura crater lake sediment sequence.
to 12 indicates an increased algal source, although some mixing of sources is still apparent. The dated macrofossil (moss) material from the two sites is associated with lower $\delta^{13}C$ values than that of organic matter in the Mt. Muhavura lake sediments, (with the exception of one sample from the Mt. Muhavura sequence at ca. 2100 cal yrs BP). This further suggests a mixed source of organic matter in the lake sediments. Further information on organic matter sources, and/or past algal productivity rates and changes in nutrient availability, may be provided by the $\delta^{13}C$ values of bulk organic matter. A change in primary productivity in the lake can affect $\delta^{13}C$ values, such that an increase in algal productivity yields a relative increase in $\delta^{13}C$, but only if supplies of dissolved carbon are limited (Meyers and Teranes, 2001; Leng and Marshall, 2004). The combined evidence of the increasing trend in $\delta^{13}C$ and increased algal input to organic matter (based on the C/N ratios) after ca. 1000 cal yrs BP suggests that algal productivity increased after this time.

7.2 Pollen data

7.2.1 Provenance and ecology of main pollen types

The main long-distance pollen types found in the sediment cores from crater sites on Mt. Muhavura and Mt. Gahinga are listed in Table 7.1 according to their relative export ability. The most common taxa in the long-distance pollen component in the pollen records from both sites are Podocarpus, Olea, Celtis, Acalypha, Urticaceae, Hagenia and Myrica.

Podocarpus, the most abundant long-distance pollen type in the cores from both sites, is a coniferous tree genus which, in western Uganda, is associated with relatively dry sites within montane forest (Hamilton, 1972; Lind and Morrison, 1974). Podocarpus pollen has a very high relative export ability, and is one of the most ubiquitous long-distance pollen types in surface samples and sediment cores at high-altitude sites in Uganda (Hamilton, 1972), and on Mt. Kenya (Ficken et al., 2002). Furthermore, Hamilton (1972) and Vincens et al. (1997) found that Podocarpus pollen was least common at sites within montane forest in which the parent taxon was present. This suggests that pollen from Podocarpus tends to be transported long distances from source, and is therefore a relatively poor indicator of the abundance of the parent taxon in local vegetation.
Table 7.1: Relative export ability of long-distance pollen enumerated in the core sequences from the Mt. Gahinga and Mt. Muhavura crater sites. Classification is after Hamilton (1972).

<table>
<thead>
<tr>
<th>Low relative export ability</th>
<th>Moderate relative export ability</th>
<th>High relative export ability</th>
<th>Very high relative export ability</th>
</tr>
</thead>
<tbody>
<tr>
<td>Araliaceae</td>
<td>Artemisia</td>
<td>Macaranga</td>
<td>Acalypha</td>
</tr>
<tr>
<td>Croton</td>
<td>Combretaceae</td>
<td>Olea</td>
<td>Celtis</td>
</tr>
<tr>
<td>Faurea</td>
<td>Dodonaea viscosa</td>
<td>Podocarpus</td>
<td>Podocarpus</td>
</tr>
<tr>
<td>Justicia</td>
<td>Galiniera</td>
<td>Urticaceae</td>
<td>Urticaceae</td>
</tr>
<tr>
<td>Maesa</td>
<td>Hagenia abyssinica</td>
<td></td>
<td></td>
</tr>
<tr>
<td>Mimulopsis</td>
<td>Helichrysum</td>
<td></td>
<td></td>
</tr>
<tr>
<td>Neoboutonia</td>
<td>Myrica</td>
<td></td>
<td></td>
</tr>
<tr>
<td>Newtonia</td>
<td>Rhus</td>
<td></td>
<td></td>
</tr>
<tr>
<td>Prunus</td>
<td>Raphanea</td>
<td></td>
<td></td>
</tr>
<tr>
<td>Solanum</td>
<td>Syzygium</td>
<td></td>
<td></td>
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<tr>
<td>Vernonia</td>
<td>Trema</td>
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Olea, the pollen of which also has a high relative export ability (Hamilton, 1972; Marchant and Taylor, 2000), is generally associated with medium-altitude moist semi-deciduous forest in eastern Africa (Hedberg, 1951). Several species of Olea commonly occur in western Uganda (Taylor et al., 1999), usually associated with relatively dry montane forest (Hamilton, 1982). Celtis spp., however, are not currently recorded in the Virunga volcanoes or surrounding Rukiga highlands (Hamilton, 1969; Owiunji et al., 2005), although seven species of Celtis occur elsewhere in Uganda, comprising a primary forest taxon in lowland and medium altitude forests (Hamilton, 1972; Taylor et al., 1999). Celtis pollen exhibits very high relative export ability; Celtis pollen in the sediment cores from the Virunga volcanoes may thus have been transported across a long distance to the site of deposition.

Acalypha pollen also has a very high relative export ability, and comprises a wind-pollinated genus of herbs and shrubs found in moist and dry lowland forest (particularly in secondary vegetation), and on disturbed land (Livingstone, 1967; Lind and Morrison, 1974). Six species of Acalypha have been recorded above 1500 m altitude in Uganda, only two of which are common over 1800 m (Hamilton, 1972), and all of which are particularly associated with edge and regenerating forest habitats, possibly due to disturbance (Lind and Morrison, 1974; Hamilton et al., 1989; Taylor et al., 1999). Urticaceae is a family of typically herbs/small shrubs and climbers found in the understory of forests. Members of the Urticaceae can be abundant in moist montane forest between ca. 2000 and 3200 m
altitude (Hamilton, 1970; 1972). Urticaceae pollen is also sometimes associated with disturbed vegetation, as some disturbance, through the creation of a more open canopy, can encourage flowering among urticaceous herbs (Hamilton and Perrott, 1980).

_Hagenia_ pollen (a pollen type of moderate relative export ability) is associated with moist high altitude forest, as _Hagenia abyssinica_ is one of the main species growing above the bamboo zone in the highest part of the montane forest belt (the _Hagenia-Hypericum_ zone) (Hedberg, 1951; Hamilton, 1972). However, this species can also be a disturbance indicator, as it is a frequent component of secondary or regenerating forest (Taylor, 1991; Jolly et al., 1997). Two species of _Myrica_ are recorded in highland Uganda: _Myrica salicifolia_, a common tree on relatively dry sites in montane forest, and _Myrica kandtiana_, a swamp forest tree, found mainly below 2000 m altitude (Hamilton, 1972; Lind and Morrison, 1974; Jolly et al., 1997). _Myrica salicifolia_ has been further described as a fire-resistant species (Finch et al., 2009). Only _Myrica salicifolia_ is currently found in the Virunga volcanoes (Owiunji et al., 2005). _Myrica_ pollen, which occurs mainly in the most recent sediments at both sites and has moderate relative export ability, is therefore assumed to be derived from _Myrica salicifolia_, and thus placed in the long-distance category.

Other long-distance pollen types occurring in the two pollen records from the Virunga crater sites include pollen from the Amaranthaceae/Chenopodiaceae families, which are generally common on recently disturbed or cultivated land, but which can also occur as forest understorey shrubs or climbers (Hamilton, 1969; 1972). _Dodonaea_ pollen, which has moderate export ability, is most common in surface samples from agricultural parts of the Rukiga highlands: _Dodonaea viscosa_ is associated with degraded and disturbed soils, and is often particularly abundant after forest clearance (Hamilton, 1982; Taylor, 1990; Taylor et al., 1999). _Rumex_ is common in agricultural parts of the Rukiga highlands (Hamilton, 1972), associated with infertile and degraded soils (Taylor, 1990). _Rumex_ is very abundant in patches of cleared vegetation in the cultivated zone on Mt. Gahinga, although the taxon has also been recorded in uncleared woodland on Mt. Gahinga (Hamilton, 1972).

The relative export abilities of the main high-altitude pollen types in the sediments from the two study sites are shown in Table 7.2. The broad ecological range represented by
Table 7.2: Relative export ability of high-altitude (Ericaceous/Afroalpine) pollen enumerated in the core sequences from the Mt. Gahinga and Mt. Muhavura crater sites. Classification is after Hamilton (1972).

<table>
<thead>
<tr>
<th>Low relative export ability</th>
<th>Moderate relative export ability</th>
<th>High relative export ability</th>
<th>Very high relative export ability</th>
</tr>
</thead>
<tbody>
<tr>
<td><em>Alchemilla</em></td>
<td><em>Anthospermum</em></td>
<td><em>Poaceae</em></td>
<td></td>
</tr>
<tr>
<td>Apiaceae</td>
<td><em>Cyperaceae</em></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Hypericum</td>
<td><em>Dendrosenecio</em></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Lobelia</td>
<td><em>Ericaceae</em> (tetrad)</td>
<td></td>
<td></td>
</tr>
<tr>
<td>Rubiaceae</td>
<td><em>Stoebe</em></td>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td><em>Typha</em></td>
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</tr>
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</table>

members of the Poaceae family, and the high relative export ability of Poaceae pollen, mean that a significant amount of Poaceae pollen may have been transported from lower to higher altitudes. *Alchemilla* pollen is often particularly abundant at high-altitude sites, and may be one of the most important local contributors of high-altitude pollen, especially given its low relative export ability (Hamilton, 1972). *Alchemilla*-scrub is currently prominent in the Afroalpine belt on the Virunga volcanoes. Ericaceae pollen is probably mostly derived from species of *Philippia* and *Erica* in the Ericaceous belt, which are common in the Virunga volcanoes, and can often produce a large quantity of local pollen (Hamilton, 1972). *Dendrosenecio* pollen is most likely to have come from within high-altitude vegetation zones, but this pollen cannot always be separated from other Asteraceae pollen, some of which can also be produced in significant quantities at lower altitudes. Flenley (1973: 134) considers *Stoebe* pollen as one of the few clear indicators of high altitude vegetation. However, this pollen type occurs only rarely in the samples from the crater sites on the Virunga volcanoes analysed in this study.

7.2.2 Synthesis and comparison of pollen records

Summary pollen diagrams, comparing the main long-distance and high-altitude taxa from the ca. 7800-year record from the crater swamp on Mt. Gahinga with the ca. 2800-year record from the crater lake on Mt. Muhavura, are shown in Fig. 7.3 and Fig. 7.4. Pollen data are plotted on a common age axis (cal yrs BP), and identical x-axis scaling is used for the percentage data from the two sites, in order to facilitate direct comparison of the records. The total long-distance/high-altitude pollen percentages, concentrations and accumulation rates for the records from both sites are also shown in Fig. 7.4.
Fig. 7.3: Comparison of long-distance pollen percentage abundances at the two sites, based on the long-distance pollen sum for each sequence. Results are plotted on a common y-axis (cal yrs BP), with identical x-axis scaling. The GPZ2/GPZ3 boundary coincides with the MPZ1/MPZ2 boundary (dating to ca. 1000 cal yrs BP). An exaggeration factor (dotted pattern) of x3 is shown for the less abundant taxa. G = Mt. Gahinga crater swamp; M = Mt. Muhavura crater lake.
Fig. 7.4: Comparison of high-altitude pollen percentages abundances from the two sites, based on the high-altitude pollen sum for each sequence. Results are plotted on a common y-axis (cal yrs BP), with identical x-axis scaling. The GPZ2/GPZ3 boundary coincides with the MPZ1/MPZ2 boundary (dating to ca. 1000 cal yrs BP). An exaggeration factor dotted pattern) of x3 is shown for the less abundant taxa. G = Mt. Gahinga crater swamp, M = Mt. Muhavura crater lake.
Information about vegetation during the period from ca. 7800 to 2900 cal yrs BP is only available in the form of pollen data from the crater swamp on Mt. Gahinga. Long-distance pollen from this period, incorporating GPZ1 and the first half of GPZ2, is particularly marked by a sustained increase in *Podocarpus* in GPZ2, along with slight increases in *Celtis* and *Olea*, and a decline in *Rapanea* and Urticaeae (Fig. 7.3). These changes suggest an increase in the extent of dry montane forest after ca. 4000 cal yrs BP in the region. High-altitude pollen is characterised by high levels of Ericaceae until ca. 5500 cal yrs BP, and a short-lived increases in *Alchemilla*, Apiaceae and Asteraceae after ca. 4500 cal yrs BP, followed by a significant rise in Poaceae (Fig. 7.4). This indicates substantial changes in high-altitude vegetation during the mid-Holocene, presumably on the Virunga volcanoes, consisting of a reduction in Ericaceous plants after ca. 5500 cal yrs BP and a later increase in Afroalpine scrub. GPZ2 overlaps with MPZ1 for the period from ca. 2800 to 990 cal yrs BP, and the pollen assemblages show many similarities, particularly for long-distance pollen types (Fig. 7.3). *Podocarpus* abundance reaches a maximum between ca. 2000 and 1000 cal yrs BP at both sites, although the increase is much more marked in the Mt. Gahinga swamp record. *Celtis*, which reaches higher relative abundance in the Mt. Gahinga swamp record than the Mt. Muhavura lake record, begins to decline from ca. 1500 cal yrs BP in both sequences (Fig. 7.3). High-altitude pollen types in the overlapping section of GPZ2 and MPZ1 show more between-site variation than long-distance pollen types: Cyperaceae and Ericaceae levels are higher in the Mt. Gahinga swamp sediments, whereas Poaceae levels are generally higher in the Mt. Muhavura lake sediments, comprising up to 55% of the high-altitude pollen sum at the latter site (Fig. 7.4). *Alchemilla* and *Dendrosenecio* are present throughout this zone in sediments from the crater lake on Mt. Muhavura, but both disappear almost completely in the record from the crater swamp on Mt. Gahinga from ca. 1500 to 1000 cal yrs BP (Fig. 7.4).

Several changes common to the long-distance pollen of both records occur in the pollen data at the GPZ2/GPZ3 and MPZ1/MPZ2 boundaries (ca. 1000 cal yrs BP). The abundance of *Podocarpus* declines significantly (and particularly abruptly in the record from Mt. Gahinga) to levels <15% of the long-distance pollen sum; *Celtis* and *Olea* levels decline, particularly in the record from Mt. Gahinga; and *Dodonaea*, *Hagenia* and *Myrica*
all increase abruptly at both sites (Fig. 7.3). Pollen from *Rumex*, Amaranthaceae/Chenopodiaceae and *Acalypha* increase towards the top of both sequences. The reduction in pollen from several montane forest tree taxa, and appearance of pollen from several herbaceous taxa associated with the recolonisation of disturbed land and with degraded soils and agricultural land, indicates a significantly reduced extent of montane forest at this time. High-altitude pollen types do not show major changes at ca. 1000 cal yrs BP (Fig. 7.4), indicating that the Ericaceous and Afroalpine vegetation occurring near the crater sites did not significantly change at this time.

A significant change apparent in the cumulative category proportions for both sites is the drop in abundances of long-distance pollen types at ca. 1000 cal yrs BP (Fig. 7.4) – influenced primarily by the fall in *Podocarpus*, which is a major component of both records between ca. 2000 and 1000 cal yrs BP, but declines abruptly after ca. 1000 cal yrs BP. Total pollen concentrations and accumulation rates are much higher in the Mt. Muhavura crater lake record than for the Mt. Gahinga crater swamp record (Fig. 7.4). Pollen concentrations in the Mt. Gahinga swamp record show a slight decline upwards throughout the record. There is a sharp drop in concentrations and accumulation rate at the MPZ1/MPZ2 boundary in the lake sediments from Mt. Muhavura, followed by a subsequent rise in values again. Pollen accumulation rates peak in the uppermost part of the records from the crater sites on both Mt. Muhavura and Mt. Gahinga.

Fig. 7.5 shows a comparison of the concentrations of the most abundant long-distance taxa from the two sites. This highlights that the lower concentrations in the Mt. Gahinga swamp sediments apply to all long-distance taxa. The decline in *Podocarpus* abundances at ca. 1000 cal yrs BP is extremely pronounced when expressed as absolute concentration values. A slight reduction in the relative amount of long-distance pollen at the MPZ1/MPZ2 boundary (Fig. 7.4) is thus most likely a reflection of the reduced concentration of *Podocarpus* pollen, rather than due to increased pollen production from, or extent of, Ericaceous and Afroalpine vegetation.
Fig. 7.5: Comparison of pollen concentrations of long-distance taxa for the sediment sequences from the crater swamp on Mt. Gahinga and the crater lake on Mt. Muhavura. Results are plotted on a common y-axis (cal yrs BP), with identical x-axis scaling. The GPZ2/GPZ3 boundary coincides with the MPZ1/MPZ2 boundary (dating to ca. 1000 cal yrs BP). G = Mt. Gahinga crater swamp; M = Mt. Muhavura crater lake.
7.3 Diatoms

Summary percentage diatom data for the sediment sequence from the crater lake on Mt. Muhavura are plotted on an age axis in Fig. 7.6. As described in Section 6.2.4, the diatom record is characterised by a combination of *Brachysira brebissonii*, *Frustulia rhomboides*, and several *Eunotia* and *Pinnularia* species. *B. brebissonii* is almost absent prior to ca. 2700 cal yrs BP, but is the most abundant taxon throughout the rest of the sequence, until it declines somewhat in the uppermost (surface) sample, dated to ca. 2003 AD. There is a diversification of *Eunotia* species after ca. 1500 cal yrs BP, and a temporary disappearance of *F. rhomboides* between ca. 800 and 100 cal yrs BP. *Gomphonema parvulum* appears only in the uppermost samples of MDZ5, which date to the 20th century. The only centric taxon to occur in the sequence, *Aulacoseira* sp., is found only in samples that post-date ca. 1000 cal yrs BP (MDZ4 and MDZ5).

The overall diatom assemblage composition of the crater lake sediments from Mt. Muhavura (i.e. the abundance of *Brachysira, Eunotia, Frustulia* and *Pinnularia* spp.) is consistent with a low pH, nutrient-poor, organic-rich shallow conditions throughout the ca. 2900-year history represented by the core sequence. The main diatom taxa are indicative of acid water of low mineral content, and occur in many mountainous lakes and peat bogs in the region. *B. brebissonii* has been found to occur in cold, acidic environments on high mountains in eastern Africa (Cholnoky, 1960), including rare occurrences in the peat bogs of Mt. Badda, in the Ethiopian highlands (Gasse, 1986). Similarly, *F. rhomboides* is known to be abundant in high-altitude aquatic ecosystems in eastern Africa, in pH conditions of below 6 (Cholnoky, 1964), and is often associated with *B. brebissonii, Eunotia* spp. and *Pinnularia* spp. (Gasse, 1986). *E. tenella*, which is the most abundant *Eunotia* species from ca. 2800 to 1500 cal yrs BP, is usually found in acidic waters and oligotrophic or dystrophic conditions (van Dam et al., 1994; Ortiz-Lerín and Cambra, 2007). In Gasse’s (1986) eastern Africa dataset, *E. tenella* was most commonly associated with subaerial habitats in the Mt. Badda peat bogs. The other *Eunotia* species occurring after the diversification of *Eunotia* taxa at ca. 1500 cal yrs BP (mostly *E. bilunaris* var. *mucophila*, *E. paludosa* var. *paludosa* and *E. rhomboidea*) are also associated with epiphytic and
Fig. 7.6: Diatom percentage data from the core sequence from the crater lake on Mt. Muhavura, plotted on an age axis (cal yrs BP). Diatoms are arranged along the x-axis according to whether they are predominantly non-planktonic or planktonic taxa (only one planktonic taxa, *Aulacoseira* sp., occurs in the core sequence). Diatom concentrations and accumulation rates are shown, along with Hill’s N2 measure of diversity.
epilithic habitats (i.e. attached to plants and rocks) in acidic waters of low mineral content (Patrick and Reimer, 1966; Gasse, 1986). *Pinnularia* species are usually associated with epipelic (i.e. attached to mud) or epiphytic habitats in shallow lakes, and with pH-neutral (e.g. *P. biceps*, Patrick and Reimer, 1966) to slightly acidic conditions (e.g. *P. microstauron* and *P. subcapitata*, Gasse, 1986).

A number of similarities in the diatom assemblages in oligotrophic montane lakes in eastern Africa have been noted by Hustedt (1949), Cholnoky (1960; 1964) and Richardson (1968), including the common association of *Brachysira, Eunotia, Frustulia* and *Pinnularia* species. Small benthic or tychoplanktonic species of *Stauroforma* (formerly classified as *Fragilaria*), sometimes in association with *Achnanthidium minutissimum*, have been recorded as abundant in some lakes on the Rwenzori Mountains (Richardson, 1968; Panizzo et al., 2008; McGlynn et al., 2010) and on Mt. Kenya (Barker et al., 2001). *Fragilaria* spp. and *A. minutissimum* are generally regarded as being tolerant of a broad range of environmental conditions (Hustedt, 1949; Richardson, 1968; Ponader and Potapova, 2007), and their low abundance in the Mt. Muhavura crater lake sediments is therefore probably not of particular ecological significance.

The shallow depth of the lake means that the majority of the water column throughout is within the photic zone, and therefore suitable for epipelic and/or epiphytic benthic taxa. The abundant aquatic macrophytes and mosses in the lake provide suitable habitat for epiphytic diatoms, while vegetation growing in the littoral zone provide a suitable subaerial habitat for *Eunotia* spp. Although *B. brebissonii* is almost absent from the lowermost sample (at ca. 2800 cal yrs BP), the abundance of *E. tenella, F. rhomboides* and *P. subcapitata* suggests occurrence of the same acidic, nutrient-poor, shallow conditions as those when *B. brebissonii* is also part of the assemblage. The slight alteration in diatoms at ca. 2400 cal yrs BP – i.e. the shift from *F. rhomboides* to *P. biceps* and *P. microstauron* – seems to represent some sort of dislocation in the community, but as these species occupy very similar benthic habitats, this change is difficult to interpret ecologically.

Diatom concentrations and accumulation rates show a peak in values at ca. 1400 cal yrs BP, and a decrease to almost zero just after ca. 900 cal yrs BP. A measure of sample
diversity (Hill’s N2) shows a peak in the same sample as that with the lowest diatom concentration (dated to ca. 880 cal yrs BP), which implies that while diatom numbers decreased at this point, the number of different species living in the lake actually increased. The low concentration is probably partly a dilution effect due to the low organic content in the sediment sample analysed for diatoms. This also coincides with the appearance of the only planktonic species to occur in the lake sediments, *Aulacoseira* sp. (possibly *A. distans*), which is usually associated with shallow, acidic waters (between pH 5 and 6) with low conductivity and low nutrient concentration (Siver and Kling, 1997).

### 7.4 Summary and comparison of the two sites

The study sites upon which the current research is largely based comprise two very different types of deposition environment: a lake of limited area and depth in a small, shallow-sided crater (on Mt. Muhavura), and a much more extensive swamp in a larger, steep-sided crater (on Mt. Gahinga). One of the benefits of analysing sedimentary records from high-altitude sites, particularly small crater sites, is that potential problems of secondary inwash of material are minimised. This is particularly the case for the crater lake on Mt. Muhavura, as the area around the lake from which inwash could occur is small, and, furthermore, the crater is not steep-sided (see Plate 4.3; page 41). Allochthonous inputs to the lake sediments (i.e. inputs not derived from within the lake) will be primarily deposited via airborne pathways, and the potential for a fluvial inwash event to deposit large quantities of locally-derived material (and the associated interpretative complications) is thus minimised. The exposed location of the site at the apex of a cone-shaped volcano – the third-highest volcano in the Virunga chain – means that airborne material carried upslope from any direction is likely to have an equal chance of being deposited at the site and incorporated into the sediment record. Airborne particle size and meteorological conditions are therefore likely to be important determinant factors in the transport and deposition of material to the lake.

The swamp in the crater of Mt. Gahinga is a larger site on a flat-topped volcano, at an altitude almost 700 m lower than that of Mt. Muhavura. The sides of the crater are steep-sided and covered in dense Afroalpine vegetation (see Plate 4.1; page 39). There may thus
be potential for erosion of material from the crater sides into the swamp. The coring site was located in the centre of the swamp, away from any visible sites of disturbance or significant inwash of material, in order to minimise this potential problem. The sediment sequence does not appear to contain any hiatuses.

The similarity of the long-distance component in the pollen records from Mt. Gahinga and Mt. Muhavura is good confirmation that they are detecting a regional signal, and not merely recording local variations. Furthermore, a more sheltered location of the crater swamp on Mt. Gahinga does not appear to have prevented deposition at the site of a significant amount of pollen from lower altitudes. Marchant (1997) argues that steep-sided catchments are less effective samplers of long-distance pollen due to topographical protection from external winds, and that pollen spectra from such catchments are generally likely to represent vegetation within the catchment. Lakes (particularly large lakes, e.g. DeBusk (1997)) are usually not as susceptible to being dominated by locally-growing taxa, when compared with swamps. This argument probably does not apply to the small lake on Mt. Muhavura, but nonetheless, locally produced pollen – in particular Cyperaceae pollen – was much more abundant in the pollen record from the Mt. Gahinga crater swamp site. The vegetation growing within the crater on Mt. Gahinga is also expected to make a greater relative contribution in the swamp than the vegetation growing adjacent to the lake on Mt. Muhavura, due to the steeper walls and the dense vegetation currently covering the steep slopes. The higher-altitude and more exposed crater lake would probably be expected both to sample a larger amount of the surrounding area and for pollen from local sources to be less abundant. Nonetheless, a comparison of the long-distance pollen records shows that both sites have detected the same regional vegetation change. The exact distance from which the long-distance pollen component has travelled is difficult to determine, but given the evidence from other studies on pollen export ability in highland eastern Africa, it seems certain that the Virunga sites are providing a good indication of environmental conditions over a wide surrounding area. The advantage of these sites is thus the fact that they are located above the main altitude of vegetation that has been substantially modified by humans. Human-modified vegetation is currently extensive at mid-altitudes (i.e. up to ca. 2200 m) in the region.
Other allochthonous inputs to the sedimentary sequences – particularly macrocharcoal and tephra – have a much higher absolute concentration in sediments from the crater lake compared with those from the crater swamp. This is mostly a dilution effect, due to the higher amount of organic material in the swamp sediments, and possibly also to differences in preservation between lake and swamp environments. The size of a charcoal fragment deposited at a site is likely to be inversely related to the distance from its source, although the effect of fire intensity – whereby more intense fires carry charcoal to a higher altitude in the atmosphere, leading to charcoal distribution over a wider area – can also be important (Patterson et al., 1987; Mooney and Tinner, 2011). Most microcharcoal quantified in the Mt. Muhavura lake sediments is nonetheless likely to be derived from a larger source area than the macrocharcoal fragments (Carcailllet, 2007; Mooney and Tinner, 2011), although quantification of this distance is impossible without detailed taphonomic studies on the production and transport of charcoal to the sedimentary basin in question.

Several particularly marked changes in different sedimentary components occur during the mid-Holocene, between ca. 5500 and 3700 cal yrs BP. A comparison of the timing of these changes is given in Fig. 7.7, showing the full age ranges based on the age models (i.e. not simply the mid-point age estimates). The occurrence of a macrocharcoal peak between the

![Fig. 7.7: Comparison of timing of major changes in several proxy measurements in the Mt. Gahinga swamp sequence during the mid-Holocene. Bars represent the full age ranges given by the age model. Age ranges for vegetation changes are based on the sample depth (analysed for pollen content) at which a change is apparent.](image-url)
two pollen samples showing high and low Ericaceae levels may be suggestive of a link between Ericaceous vegetation and fire activity. The major tephra peak is shown to be a separate event to the macrocharcoal peak, suggesting that the fires represented by the macrocharcoal peak at ca. 5200 cal yrs BP were not linked to volcanic activity.

The increase in dry montane forest taxa during the mid-Holocene (i.e. increase in *Olea* followed by rise in *Podocarpus*) does not occur as one synchronous event, but is spread over a period of up to 500 years, suggesting a gradual vegetation change at this time. A second, smaller macrocharcoal peak overlaps with the initial increase in *Olea* and *Alchemilla* pollen; a possible relationship between vegetation and fire dynamics during the mid-Holocene will be discussed in Chapter 8.

Fig. 7.8 shows the full age ranges of the major changes in pollen, charcoal and tephra in the sedimentary records from both sites between ca. 1000 and 700 cal yrs BP. The marked change in pollen content at the GPZ2/GPZ3 and MPZ1/MPZ2 boundaries (the

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**Mt. Muhavura crater lake**

*Podocarpus* decline

Tephra peak

Macrocharcoal peak

**Mt. Gahinga crater swamp**

*Podocarpus* decline

Tephra peak

Macrocharcoal peak

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**Fig. 7.8:** Figure to highlight the age ranges associated with changes in the pollen, tephra and macrocharcoal records between ca. 1000 and 700 cal yrs BP from the two study sites. Bars represent the age range given by the age models. Lines indicate overlapping age ranges for the same feature at the two sites. Calculation of the age range for the *Podocarpus* decline was based on the sample depth at which *Podocarpus* pollen sharply decreases (86-87 cm for Mt. Muhavura crater lake; 270-271 cm for Mt. Gahinga crater swamp). The age ranges for the tephra and macrocharcoal peaks were calculated based on the depth range over which they occurred.
Podocarpus decline) occurs, according to the Mt. Gahinga record, between 940 and 890 cal yrs BP – an interval that is within the range of the Mt. Muhavura estimate (Fig. 7.8). Furthermore, the clear separation of the timing of the Podocarpus decline and the tephra and macrocharcoal peaks shown by the Mt. Gahinga sequence suggests that the broad, overlapping age ranges given by the age model for the changes in the Mt. Muhavura sequence are masking the discrete nature of the Podocarpus decline from the other two events. The date estimates for the tephra and charcoal peaks at the two sites are slightly offset, but this may merely be a factor of the sampling resolution used. The sedimentary records from the two sites appear to have recorded the same regional signals, rather than non-synchronous, and more localised events. The best estimate of the date of the Podocarpus decline is between ca. 940 and 900 cal yrs BP – slightly after the zone boundaries at ca. 1000 cal yrs BP – hereafter referred to as an estimate of ca. 900 cal yrs BP. The similarity of timing (despite a slight apparent offset) of the largest relative macrocharcoal peaks in the two records is an indication that fire events may have affected both sites at around the same time. The implications of these features will be further discussed in Chapter 8.
Chapter 8 Discussion

This chapter provides a discussion of the results presented in chapters 5–7, in the context of this study's research questions. As outlined in Chapter 1, this thesis aims to reconstruct the environmental history of the Albertine Rift during the Holocene, based on high-altitude sediment records from two of the Virunga volcanoes. In particular, the research aims to (i) examine the extent to which high-altitude sites have been affected by monsoanally-driven climate change during the mid- to late Holocene; (ii) evaluate whether high-altitude sediment records show evidence for increased human environmental impact in recent millennia; and (iii) provide a perspective on the roles of climate change, fire, volcanic activity and human activity as drivers of environmental change at high-altitude sites during the Holocene. These three research questions will be addressed in Section 8.1, and will be followed (in Section 8.2) by a discussion of the wider implications of these findings.

8.1 Holocene environmental change in the Albertine Rift

8.1.1 Mid-Holocene climate change

Sedimentary evidence from the crater swamp on Mt. Gahinga highlights a number of changes during the mid-Holocene, many of which are corroborated by the bottom part of the sequence from the crater lake on Mt. Muhavura. Vegetation changes indicated by the pollen data suggest that both the distribution of high-altitude vegetation belts and the composition and extent of lower montane forest experienced marked changes during the mid-Holocene.

The fall in Ericaceae pollen at ca. 5000 cal yrs BP, and the sustained low levels of Ericaceae pollen throughout the rest of the record (not only in swamp sediments from the site on Mt. Gahinga but also throughout the 2800-year record from the lake on Mt. Muhavura), indicates a significant (and permanent) reduction in the relative extent of Ericaceous vegetation during the mid-Holocene. The increase in pollen from Afroalpine taxa (including Alchemilla, Asteraceae, Dendrosenecio and Poaceae) after ca. 4500 cal yrs BP suggests the expansion of open Afroalpine vegetation in replacement of Ericaceous vegetation at high altitudes on Mt. Gahinga. At present, there is no well-developed
Ericaceous zone on Mt. Gahinga, but Afroalpine vegetation occurs from ca. 3300 m to the summit (3474 m). This is a lower altitude than usual for the occurrence of well-developed Afroalpine vegetation: the upper limit of Ericaceous belt vegetation usually occurs between an altitude of 3500 and 4100 m (Hedberg, 1951). The pollen record from Mt. Gahinga provides evidence that Ericaceous vegetation formerly occurred on the volcano during the early Holocene, and that the replacement of Ericaceous vegetation with Afroalpine vegetation occurred during the mid-Holocene. The earlier decline in Ericaceae (compared with Afroalpine vegetation increase) suggests that this was not rapid encroachment of Afroalpine vegetation into an area occupied by Ericaceous vegetation, but rather indicates an externally-driven factor for the Ericaceae decline, followed by the later downward expansion (or colonisation) of Afroalpine vegetation. This indicates a downward retreat of the upper Ericaceous belt limit. A brief expansion of upper montane forest (*Hagenia*) does not occur until after ca. 3500 cal yrs BP, suggesting that upward expansion of montane forest vegetation was not a factor in the Ericaceae decline.

Long-distance pollen evidence from the Mt. Gahinga crater swamp sediments indicates an expansion of dry montane forest types from ca. 4100 cal yrs BP, consisting of an initial increase in *Olea* followed by a rise in *Podocarpus* from ca. 3800 cal yrs BP. An increase in drier forest types is further supported by a decrease in moist forest undergrowth from ca. 4100 cal yrs BP. This dry montane forest vegetation is likely to have covered a wide landscape surrounding the Virunga volcanoes, as much of the surrounding area is at altitudes that are likely to have supported montane forest, prior to widespread human activity – particularly the upper slopes of the undulating ridge-and-valley topography, suitable for drought-tolerant forest. These changes occur at the same time as the expansion of Afroalpine vegetation at high altitudes, suggesting that vegetation at different altitudes is responding to a common forcing factor – in this case, an increasingly dry climate. However, the offset of at least 1000 years between the fall in Ericaceae and the expansion of Poaceae and *Podocarpus* further suggests that these vegetation changes are not responding to a single, abrupt climate event.

Macrocharcoal evidence shows that the fall in Ericaceae pollen may have been linked to increased fire activity. Ericaceous vegetation often burns extensively during particularly
dry seasons (Hedberg, 1951), which suggests that an increased seasonality of climate, including a longer dry season allowing for the accumulation of a greater amount of combustible material, was a factor in causing the changes in high-altitude vegetation at this time. There is a further, smaller peak in macrocharcoal occurring at ca. 4100 cal yrs BP, coinciding more closely with the beginning of the increase in Afroalpine vegetation – which may indicate that fires were continuing to suppress the re-establishment and re-expansion of Ericaceae vegetation. These changes in climate and fire dynamics during the mid-Holocene at high altitudes are likely to have also affected vegetation on the five other main peaks of the Virunga volcanoes (Mts. Muhavura, Sabinyo, Bisoke, Karisimbi and Mikenro), all of which reach a higher altitude than Mt. Gahinga, and support more extensive areas of Ericaceae and Afroalpine vegetation. If Ericaceae vegetation was more extensive on the lower-altitude Mt. Gahinga in the early Holocene, then this is also likely to have been the case on the other volcanoes. Furthermore, the persistence of low levels of Ericaceae pollen in the sediment sequence from Mt. Gahinga, even those that have accumulated recently, indicates that Ericaceae pollen from other volcanoes is being incorporated into the sedimentary record at the crater swamp on Mt. Gahinga (as such vegetation is not widespread on Mt. Gahinga today). This provides evidence that the high-altitude pollen sediment record from the crater swamp on Mt. Gahinga is representative of the high-altitude zones across the Virunga volcanoes, not just the volcano from which the sediment core was obtained.

An increase in aridity in the mid-Holocene is a common feature of tropical African palaeoclimate records, linked to an overall precession-driven weakening of the monsoonal system causing increased aridity during the late Holocene (Marchant and Hooghiemstra, 2004), although a particularly abrupt drying event seems to have occurred near 4000 cal yrs BP (Tierney et al., 2011). However, within tropical eastern Africa, and even within the Albertine Rift itself, the climate response to insolation changes in different areas has been complex. The transition from humid to arid conditions was not widely synchronous, but began in some areas >5000 cal yrs BP, while not occurring in other areas until almost 3000 cal yrs BP (Russell et al., 2003c; Russell and Johnson, 2005). The ice core record from Mt. Kilimanjaro shows a layer of dust deposited at 4000 cal yrs BP, which is thought to reflect
a regional aridity peak (Thompson et al., 2002). Several lake records indicate drought conditions at around 4000 cal yrs BP, although the timing of lake-level falls and desiccations varies, and most sites show evidence for more than one drought phase during the late Holocene (e.g. Barker et al., 2002; Russell et al., 2003a; Stager et al., 2003; Russell and Johnston, 2007). In particular, sedimentary evidence from Lake Edward (one of the lakes of the Albertine Rift, located ca. 80 km north of the Virunga volcanoes) indicates the onset of drier climate at 5200 cal yrs BP (Russell et al., 2003a). This coincides with the macrocharcoal peak in the Mt. Gahinga sediments – providing evidence that the changes at high altitude evident in the sedimentary record from Mt. Gahinga were also felt elsewhere in the Albertine Rift. This further substantiates the argument that increasing aridity affected the fire regime in the Virunga volcanoes, causing substantial changes to high-altitude vegetation, particularly fire-susceptible Ericaceous vegetation. The changes in montane forest vegetation (i.e. the *Podocarpus*-rise) are also indicative of a response to climate aridification, whereby a relative drying of soil conditions, particularly on exposed slopes in the ridge-and-valley topography, would provide conditions suitable for the growth of *Podocarpus* and *Olea* (Jolly et al., 1997). The complexity of vegetation responses to mid-Holocene climate change is highlighted further by a comparison of the records of vegetation change from the Virunga volcanoes with those from other sites in high-altitude eastern Africa. The timing of vegetation change varies slightly from site to site, but a broadly synchronous pattern is apparent, reflecting the wide-reaching influence of climate as a driver of vegetation change during the mid-Holocene. The pollen record from Muchoya Swamp (2260 m altitude) in southwest Uganda, located ca. 20 km northeast of the Virunga volcanoes, records a significant change in forest taxa at ca. 3700 cal yrs BP, comprising a major increase in *Podocarpus* pollen and decline in Urticaceae pollen (Taylor, 1990; 1992) – occurring at the same time as that observed in the Mt. Gahinga record. At Ahakagyezi Swamp (1830 m altitude), also located in southwest Uganda (ca. 40 km northeast of the Virunga volcanoes), the *Podocarpus*-rise is dated to ca. 4300 cal yrs BP (Hamilton et al., 1989; Taylor, 1993), slightly earlier than the new evidence from the Virunga volcanoes suggests. This also coincides with a major expansion of *Syzygium* swamp forest at Ahakagyezi Swamp, which suggests a drier swamp
surface (Taylor, 1990; 1993). Mubwindi Swamp (2100 m altitude) in Bwindi Impenetrable Forest has a sedimentary hiatus through much of the Holocene, which is thought to have been caused by erosion due to a wetter climate (Marchant et al., 1997). By the late Holocene (re-sedimentation is established by ca. 2100 cal yrs BP), *Podocarpus* and *Olea* were common components of moist lower montane forest at Mubwindi Swamp (Marchant and Taylor, 1998). Similarly, several sites in the Ethiopian highlands contain a hiatus in peat accumulation during the early to mid-Holocene, linked to a wetter climate (Mohammed and Bonnefille, 1998). Records from two sites located in the Afroalpine belt on Mt. Elgon (Lake Kimilili, 4150 m altitude, and Koitoboss Bog, 3940 m altitude) show a dramatic increase in *Podocarpus* pollen, along with decreases in Urticaceae, *Rapanea, Macaranga* and *Hagenia*, at ca. 4100 cal yrs BP (Hamilton, 1982). A similar rise in *Podocarpus* occurs at ca. 4000 cal yrs BP in pollen records from Lake Mahoma (2960 m altitude) and Lake Kitandara (3990 m altitude) in the Rwenzori Mountains (Livingstone, 1967), although the radiocarbon dating for these sites is not very secure (Hamilton, 1982). Lake Rutundu (3078 m altitude) on Mt. Kenya also shows an increase in *Podocarpus* pollen and decline in Ericaceae from ca. 4000 cal yrs BP (Wooller et al., 2003). In the Bale Mountains of Ethiopia, a significant rise in *Podocarpus* and other dry montane taxa (*Juniperus* and *Olea*) occurs from ca. 4500 cal yrs BP (Umer et al., 2007). This pollen record also shows a sharp decline in Ericaceae pollen at ca. 4500 cal yrs BP, interpreted as a downward retreat of the upper altitudinal limit of the Ericaceous belt in the Bale Mountains (Umer et al., 2007). The expansion of upper montane forest at the same time implies that Ericaceous vegetation was much reduced in extent during the late Holocene compared with the early Holocene (Umer et al., 2007). Ericaceae pollen is also common in sediments from Muchoya Swamp, dating to the early Holocene, although at this site some of this pollen may be from *Erica* plants growing on drier parts of the swamp, as is the case today (Taylor, 1990).

The principal cause of the changes in high-altitude and montane forest vegetation during the mid-Holocene can thus be interpreted as primarily a response to a drier climate on a large scale, affecting the entire eastern African region, related largely to the precession-induced weakening of the monsoon system at the end of the African Humid Period.
Apparently non-synchronous changes in high-altitude and montane forest vegetation can be attributed to the differential effect of climate on different vegetation types, and may also be related to the occurrence of multiple drought intervals during the mid- to late Holocene. Pollen data from the crater swamp on Mt. Gahinga provide further confirmation of the ubiquity of the Podocarpus-rise across montane sites in eastern Africa at ca. 4000 cal yrs BP. Furthermore, vegetation changes occurring at high-altitudes indicate an earlier response to climate change than that associated with the Podocarpus-rise, and highlights the close links between climate (particularly seasonality) and fire regimes.

8.1.2 Late Holocene environmental change: the influence of increased human activity

The first unequivocal evidence for human impact detected by the sedimentary records from the Virunga volcanoes dates to within the last thousand years. Prior to this, between ca. 4000 and 1000 cal yrs BP, the vegetation around the Virunga volcanoes continued to be largely comprised of lower montane forest (established during the mid-Holocene at the Podocarpus-rise). This vegetation type expanded further during this period: Podocarpus levels reach a maximum abundance (both relative and absolute) between ca. 1500 and 1000 cal yrs BP, although some montane forest taxa begin to decrease at this time (e.g. Celtis spp.). The vegetation changes centred on ca. 900 cal yrs BP at both study sites are indicative of the first marked onset of human activity. Pollen evidence indicates an abrupt and widespread reduction in lower montane forest extent and replacement with herbaceous disturbance-indicator taxa, associated with both the recolonisation of disturbed land and with degraded soils and agricultural land. These changes are consistent with widespread clearance of lower montane forest for cultivation, thus invoking human agency as the primary cause. Furthermore, no major changes in high-altitude vegetation occurred at ca. 900 cal yrs BP. As climate changes would be expected to affect high-altitude vegetation as well as lower montane vegetation (as occurred during the mid-Holocene), climate can thus be ruled out here as a major factor in causing the changes in montane forest vegetation. This is not to say that climate did not experience any fluctuations during this period – extensive evidence from other sites in the Albertine Rift, and elsewhere in tropical eastern and central Africa, indicates a drought period from ca. 950 to 750 cal yrs BP (Verschuren et al., 2000; Alin and Cohen, 2003; Stager et al., 2003; Russell and Johnson, 2005; Ryner
et al., 2008; Ryves et al., 2011) – but that human agency at this time exercised a greater influence on vegetation change than climate. Nonetheless, a drier climate may have been a factor in causing some of the other sedimentary changes observed at this time, and furthermore, changes in the patterns of human-environment interaction occurring at this time may have been influenced by the onset of a drought period.

With such an apparently major episode of forest clearance, an increase in fire occurrence (due to clearance of the forest by burning) might be expected to occur. Microcharcoal levels do not suggest a major burning episode on a widespread scale, which is likely to reflect the different provenance areas represented by pollen and microcharcoal particles: *Podocarpus* pollen, in particular, has a very high relative export ability, and is likely to derive from a relatively large source area when compared with microcharcoal particles. An increase in the frequency and/or intensity of vegetation fires locally, indicated by macrocharcoal peaks in the sediment records from both Mt. Gahinga and Mt. Muhavura, occurs at ca. 800 cal yrs BP, which is ca. 100 years after the onset of forest clearance. This increase in localised fires in the Virunga volcanoes, along with the continued decline in montane forest pollen taxa, suggests that by ca. 800 cal yrs BP, forest clearance occurred in closer proximity to the volcanoes than at ca. 900 cal yrs BP. However, the Mt. Muhavura record indicates that increased macrocharcoal occurrence may have occurred from as early as ca. 950 cal yrs BP. Separating the timing of regional forest clearance and localised fires is therefore difficult, and deserving of further research. Regardless of whether peak macrocharcoal concentrations occurred at the same time as montane forest pollen decline, or slightly afterwards, the conclusion stands that major forest clearance on both a regional and local scale occurred on the Virunga volcanoes during the past millennium, beginning at ca. 900–800 cal yrs BP.

A further dimension to the influence of human activity on vegetation change during the last millennium is provided by the fluctuations in *Hagenia* pollen, which in this setting is mostly likely indicative of a change in the composition of upper montane forest: *Hagenia abyssinica* is an important component of upper montane forest on the Virunga volcanoes at present, yet *Hagenia* pollen features prominently in the palynological record only since ca. 900 cal yrs BP. *Hagenia* can also be found in regenerating forest at lower altitudes, but is
most commonly associated with forest at the upper tree line (Jolly et al., 1997). The increased occurrence of *Hagenia* after ca. 900 cal yrs BP may also be linked with the change in fire dynamics suggested by the major macrocharcoal peak occurring at this time, as *Hagenia abyssinica* is sometimes characterised as a fire-tolerant species (Rucina et al., 2009). *Myrica salicifolia*, from which the increased *Myrica* pollen from ca. 900 cal yrs BP onwards is likely to be derived, is also a fire-tolerant species (of dry montane forest). The increased occurrence of fire-tolerant taxa may reflect a vegetation response to increased localised burning, as indicated by the major macrocharcoal peak occurring at this time. Furthermore, the apparent change in the composition of upper montane forest at ca. 900 cal yrs BP, coincident with forest clearance and increased disturbance indicators, suggests that the influence of human societies on montane forest may be more pervasive – and extend to higher altitudes – than previously considered. The palaeoecological evidence from the crater sites on the Virunga volcanoes thus shows the extent to which human activity over the past millennium has altered not just the extent of forest, but also its composition.

The vegetation changes occurring at ca. 900 cal yrs BP coincide with the approximate timing of Late Iron Age onset in Rwanda and Uganda, which was associated with widespread social, political and economic changes (Sutton, 1993; Ashley, 2010). These changes included the development of large centralised settlements with new ceramic cultures in western Uganda (Robertshaw, 1994; Ashley, 2010), and possibly the development of increasingly specialised agricultural practices, such as specialised banana farming on permanent plantations (Schoenbrun, 1993b). Direct archaeological evidence for agriculture from several sites in Rwanda indicates cultivation of millet and sorghum by ca. 1500 cal yrs BP, although for cave sites adjacent to the Virunga volcanoes, evidence for occupation dates to ca. 1000 cal yrs BP (Giblin and Fuller, 2011). Existing palaeo-ecological evidence from the Albertine Rift and adjacent areas yields different patterns and timing of human-induced vegetation change over the past two millennia, with some sites showing localised vegetation change dating to as early as ca. 2300 cal yrs BP, which may be associated with human activity (e.g. Muchoya Swamp; Taylor (1990), and also Kabata Swamp; Taylor et al. (1999)). However, the extent of human influence on vegetation does not appear to have been widespread until ca. 900 cal yrs BP (e.g. Ssemmanda et al., 2005;
Russell et al., 2009). The evidence from the Virunga volcanoes presented in this study further attests to this, as widespread forest clearance linked to human activity is not indicated until ca. 900 cal yrs BP.

The records from the crater sites on the Virunga volcanoes are expected to capture a more regional vegetation signal than sites at lower altitudes, due to both the tendency for pollen to be transported from low to high altitudes and to the lack of direct human influence on the immediate catchments of the crater sites. Dates for forest clearance prior to ca. 900 cal yrs BP, such as that from Muchoya Swamp, are thus likely to reflect a localised clearance close to the site. The drastic change in vegetation detected in the crater sites from the Virunga volcanoes, and the similar features observed in both sites, is perhaps the best indication obtained to date of the true extent of forest clearance in the region around the beginning of the second millennium AD. Furthermore, the abundance of macrocharcoal, and particularly the changes in the composition of upper montane forest (i.e. expansion of *Hagenia*), shows that the extent of human modification of vegetation has been pervasive at all altitudes within the montane forest belt, even in areas currently gazetted within national parks and viewed as comprising pristine natural vegetation. Similar findings have emerged from tropical forests in other areas (e.g. the Amazon basin), where evidence from archaeological and palaeoecological investigations suggests that some so-called virgin forests may not be as pristine as originally thought, but may have already undergone substantial modifications due to the influence of past human activity (Willis et al., 2004; Mann, 2008).

Although there is no evidence for major changes in Afroalpine or Ericaceous vegetation distribution during the late Holocene, evidence from other proxy measurements on sediments from the crater lake on Mt. Muhavura (particularly diatoms, tephra and geochemistry) suggests that environmental changes occurred during this time at this high-altitude site, which may be related to regional climate changes. Diatom data – based on the expansion and diversification of *Eunotia* spp. – suggests a possible expansion of subaerial habitats around the lake’s margins at ca. 1500 cal yrs BP, which would be consistent with a slight lowering of lake level. The lake’s shallowness and small size, in addition to its enclosed nature in a small crater basin, mean that even a small drop in lake level would
have a significant effect on the shallow lake margins by exposing areas of the littoral zone. A major drop in lake level would be likely to lead to near or total desiccation of the lake – a scenario for which there is no direct evidence in the sedimentary sequence. The expansion and diversification of *Eunotia* spp. do not coincide with any other sedimentary changes in the sequence, and thus no obvious causes of the change are apparent – except a response to lowered lake level, which may be related in turn to a drought episode. Other sites in the Albertine Rift – particularly the record from Lake Edward – attest to pronounced drought between 1540 and 1460 cal yrs BP (Russell and Johnson, 2005), and diatoms preserved in sediments from the crater lake on Mt. Muhavura may thus be detecting this climate signal.

Between ca. 900 and 800 cal yrs BP, a significant (albeit very brief) drop in diatom concentration and accumulation rates occurs, coinciding with the occurrence of a marked peak in sediment minerogenic content, and with the maximum abundance of tephra and macrocharcoal. No major changes in diatom assemblage composition occur, although the subsequent occurrence of *Aulacoseira* sp., which are poor competitors for silica and tend to expand during periods of low diatom production (Wolfe et al., 2000; Leira, 2005), may reflect a slight increase in silica to the lake system, provided by the deposition of tephra. This evidence, together with the existence of a charcoal peak, may indicate that regional aridity between ca. 950 and 750 cal yrs BP (as noted earlier) affected the lake at this time: increased erosion and soil movement due to a drier climate may have caused the observed changes in the diatom record. Additional evidence of changes within the lake system is provided by organic geochemistry (C/N ratios and δ¹³C). This evidence indicates that the composition of organic matter in the lake after ca. 900 cal yrs BP was characterised by increased input from algal sources (possibly related to higher algal productivity), and a decline in the incorporation of material from aquatic macrophytes and/or terrestrial plants. Furthermore, pollen from aquatic plants (primarily *Typha*) decreases after ca. 900 cal yrs BP, indicating that such vegetation may have been an important contributor to organic matter formation in the lake prior to this time.

Overall, a picture emerges of a perturbation to the crater lake on Mt. Muhavura between ca. 900 and 800 cal yrs BP, involving deposition of tephra and macrocharcoal, a brief but
abrupt decline in diatom concentration, and a reduction of aquatic vegetation, which may have been at least partly related to an increase in aridity noted in the region at this time. Following this event, aquatic vegetation does not recover to the same extent as previously, and a slight shift occurs in the diatom assemblage composition, although aquatic conditions (low acidity, low mineral content, shallow water) remain relatively unchanged. The timing coincides with clearance of lower montane forest and localised fires, described earlier in this section. The changes evident within the crater lake suggest a response to tephra and/or charcoal deposition, including an effect on the extent of aquatic vegetation. The amount of tephra deposited was very low, but in a silica-poor, organic-rich system, this can be sufficient to influence diatom communities (P. Barker, pers. comm.) – in this case, facilitating the expansion of species that are relatively poor competitors for silica. A change in lake-level is not directly evident at this time, but rather these changes indicate the response of an isolated aquatic ecosystem to perturbations in the system, in the form of changes in the influx of sedimentary components.

8.1.3 Relative driving factors of environmental change

This study highlights the complexity of factors influencing the environment of the Virunga volcanoes and adjacent parts of the Albertine Rift during the Holocene, as recorded in the crater sediments from Mt. Gahinga and Mt. Muhavura, and presents evidence that climate, fire, human activity and volcanic activity have all acted as important (and sometimes interacting) factors influencing environmental change. During the mid-Holocene, climate was the most important factor affecting environmental change, as manifested by the vegetation changes at both high and lower altitudes, all of which are consistent with a shift towards drier climatic conditions. Climate-fire dynamics were also important at this time, particularly with regard to seasonality and length of dry season: a rapid initial response of high-altitude vegetation to mid-Holocene aridity at ca. 5000 cal yrs BP occurred due to the susceptibility of Ericaceous vegetation to fire during long dry seasons. Lower montane forest, and some high-altitude communities, does not show major changes until about a thousand years later, which highlights the differing sensitivities of vegetation to climate change, and may also be related to the complexity of climate change during the mid- to late Holocene, which was not a single aridification event but rather a gradual and multi-step process of climate aridification.
Humans have played an increasingly prominent role in shaping landscape during the late Holocene. This study shows that human activity has been an important component of environmental change during the past millennium, both directly and indirectly. The direct effects of human activity are interpreted as the clearance of forest, most likely by slash-and-burn clearance, and replacement of the previous forest cover with cultivated land or secondary forest. This is shown by the widespread decline in *Podocarpus* at ca. 900 cal yrs BP, and subsequent changes in vegetation. Indirect environmental effects may arise where human activity leads to suitable conditions for vegetation types that were not previously widespread. The alteration of fire dynamics by humans is an important dimension of such indirect effects. In the current study, this is shown by an increase in *Hagenia* after ca. 900 cal yrs BP, along with increases in other fire-prone taxa such as *Myrica*, indicating a response to human-induced vegetation clearance and increased fire occurrence that affected not only lower montane forest where the clearance was taking place, but also the composition of upper montane forest.

The influence of volcanism on environmental conditions during the Holocene is perhaps more difficult to separate from other potential drivers of change. The lack of any major, visible tephra horizons in the sediment sequences suggests that volcanic activity may not have occurred close enough to the study sites to act as a major influence on vegetation (e.g. via the effect of lava flows or ignition of fires). However, the amount of tephra deposited in the sediments on occasion was sufficient to have an effect on aquatic conditions and on the assemblage composition of silica-using algae (diatoms) in the lake, and may have had a fertiliser effect on terrestrial vegetation. A large macrocharcoal peak, amongst other sedimentary changes (including increased overall minerogenic content), occurs at the same time as the most significant tephra peak at ca. 900 cal yrs BP. While these do not necessarily reflect a related cause (as microscopic tephra shards and macrocharcoal may not necessarily represent the same spatial scale of environmental change), their occurrence at this time, which also coincides with a drought period indicted by several palaeoclimate records in the region (e.g. Russell and Johnson, 2005; Ryves et al., 2011), may reflect an increase in local fires caused by the combination of volcanic activity and the accumulation of dry biomass due to a more arid climate. Nonetheless, the perturbations occurring in the
lake at ca. 900 cal yrs BP are coincident with significant human-induced vegetation change – thus pointing towards inter-related causes of direct human impact, human-induced alteration of fire dynamics, climate change, and volcanic activity, all of which influenced different aspects of environmental change at ca. 900 cal yrs BP.

8.2 Broader implications of the research

The sedimentary records presented in this study indicate the increasing importance of human activity during the late Holocene as a major component of environmental change. The widespread effect of human activity on the landscape of the Albertine Rift as far back as ca. 900 cal yrs BP raises some pertinent issues, and highlights the long duration of interaction between humans and their environment since long before modern times.

One obvious question that arises from the current research is why humans only began to impact environmental conditions in a tangible way within the last millennium, given the long presence of humans on the African continent, and notably eastern Africa, and the abundant resources that would have been present locally. Evidence for Early Iron Age occupation in the Albertine Rift and extending eastwards to the area around Lake Victoria is widespread, in the form of the Urewe ceramic tradition (dated to ca. 2500 to 1200 cal yrs BP) (see Ashley (2010) for a recent review). Schoenbrun (1998), based on historical linguistic evidence, postulates that the main population element in the area at this time were Bantu-speakers, growing crops suited to relatively moist forest environments and utilising iron tools, and thus taking advantage of the moist forest environments in the Albertine Rift as suitable sites for their food-producing economy. The Early Iron Age marks a significant change in the societal history of the region, and Schoenbrun (1993a) suggests that climate aridification during the late Holocene, and the associated reduction of moist forest vegetation in many areas (Taylor, 1992; Taylor et al., 1999; Vincens et al., 2003) may have necessitated a diversification on the part of Bantu-speakers into environments other than moist forest in order to increase food security. This, in turn, may have further intensified the pressure on forest land, possibly accelerating the forest decline, and accounting for the successful expansion of Bantu-speaking peoples during the Early Iron Age (Schoenbrun, 1993a). However, there is no evidence for significant human-
induced forest clearance in the sedimentary records from the Virunga volcanoes until ca.
900 cal yrs BP, although some Early Iron Age archaeological sites have been identified in
the surrounding area (Van Nooten, 1983; Giblin, 2008; 2010). This most likely indicates
that human modification of the landscape prior to this time was low-intensity and
temporary, and thus can be detected in sedimentary records only when the sediment basin
is proximal to former areas of settlement. The records from the Virunga volcanoes, located
at a higher altitude than would have been directly affected by human-induced vegetation
clearance, do not detect any Early Iron Age settlement signal, confirming the low intensity
(and possible mobility) of settlement in the surrounding area at this time.

Distinguishing between the roles of human agency and climate as factors in influencing
vegetation change can be challenging, especially as there are often similarities of timing
between societal and environmental (including climate) changes. The development of
theories expounded to explain human–environment inter-relationships is characterised by a
shift, largely in the twentieth century, from notions of a mainly unidirectional human–
environment interaction, leaning towards deterministic explanations of societal change,
towards recognition of the mutual interaction of people with their physical environment
(Moran, 2000; 2005). In a summary study of climate and societal history in eastern and
southern Africa, Holmgren and Öberg (2006) conclude that societal changes often (but not
always) coincide with climatic changes, and that climate change can be a particularly
important trigger of either societal catastrophe or of new developments if it coincides with
times of socio-economic and political instability. Significant climate changes – such as
increasing aridity during the late Holocene – have a tangible effect on various aspects of
the natural environment, including vegetation cover and soil moisture. However, human
activity can also be an important contributing factor to environmental conditions (by
affecting, for example, soil conditions, the composition of vegetation, water and
atmosphere quality), including in habitats previously thought pristine (Willis et al., 2004).
In the context of human–environment interactions, this highlights that while a purely
deterministic relationship between climate and societal change is unlikely, climatic change
can nonetheless play an important role in influencing human society.
A direct climatic cause of the vegetation changes that occur at ca. 900 cal yrs BP in the records from the Virunga volcanoes is unlikely, as outlined in Section 8.1.2. However, the possible indirect role of climate change cannot be discounted in the analysis of vegetation changes during the late Holocene, even those associated with human-induced vegetation clearance at ca. 900 cal yrs BP. Several pronounced drought phases during the past millennium are evident from lake-level records, which may have had an appreciable impact on human societies by influencing the way in which societies interacted with their environment. Major changes in settlement patterns and political structure in the interlacustrine region during the Late Iron Age may be related to increased risk buffering due to drought phases: Taylor et al. (2000) concluded that the establishment of centralised authority during the Late Iron Age could have been an attempt to buffer against risks of food shortages. Many areas experienced drought conditions between 950 and 750 cal yrs BP (Verschuren et al., 2000; Stager et al., 2003; Stager et al., 2005; Russell and Johnson, 2005; Russell et al., 2007), supporting the suggestion of an increased need for risk buffering, indicating that climatic change may have been an important contributory factor in the societal changes that resulted in the emergence of large nucleated settlements and political centralisation throughout much of central and eastern Africa.

A further issue arising from the sedimentary records from the Virunga volcanoes relates to the importance of past human activity in shaping the modern landscape of the Albertine Rift. In other parts of the world (e.g. China), a shift from “nature-dominated” to “human-dominated” landscapes (Dearing, 2008: 3) has been observed as having occurred as early as ca. 4800 cal yrs BP. However, in the Albertine Rift, such a marked increased role of human activity as a cause of widespread environmental change is not apparent during the mid-Holocene, but is very much apparent from ca. 900 cal yrs BP onwards, when human modification of the landscape largely eclipses the role of other components, such as climate variability. Such findings have implications on the concept of the “Anthropocene”, a term recently applied to the past ca. 200 years in light of the escalated effects of humans on the global environment since that time (Crutzen, 2002: 23), and which has gained much coverage in both scientific and media outlets (e.g. Ruddiman, 2003; Dalby, 2007; Steffen et al., 2007; Kolbert, 2011; Koster, 2011). The boundary is placed in the late 18th century
due to the clear acceleration in concentrations of greenhouse gases dating to the Industrial Revolution, and the subsequent global environmental effects of a tenfold rise in human populations since that time (Crutzen, 2002; Crutzen and Steffen, 2003). However, a broader interpretation of the Anthropocene can be viewed in terms of the timing of widespread human-induced landscape modification – in which case, the evidence from the Virunga volcanoes suggests that the Anthropocene has been occurring in the Albertine Rift since ca. 900 cal yrs BP.

Evidence that human activity has extensively modified the landscape of the Albertine Rift since ca. 900 cal yrs BP, including areas at high altitude on the Virunga volcanoes that are currently gazetted as national parks, have significant implications for conservation planning. The idea of preserving the last remnants of a natural landscape untouched by humans is a powerful justification for international conservation agendas in the Albertine Rift, with Bwindi Impenetrable Forest and the three parks of the Virunga volcanoes viewed as the last “islands of forest” remaining of a formerly “extensive forest massif” (Rainer et al., 2003: 190). Historically, conservation measures have followed a protectionist approach to conservation planning (often known as a “fences and fines” or “fortress conservation” approach) (Adams and Infield, 2003: 177). More recent approaches (which have received increased international attention over the past decade – see e.g. Chapin (2004) and Brockington et al. (2006)) recognise the importance of a more inclusive approach that incorporates the needs of local communities into conservation planning, rather than viewing local populations in terms of a “prime threat to the protection of these forest blocks” (Rainer et al., 2003: 192). Balancing the needs of local communities against the high-cost (and high-income), intense conservation focused largely on the critically endangered flagship species, the mountain gorilla (e.g. Robbins et al., 2011), remains a challenge. The collaborative aims of the International Gorilla Conservation Programme, and the agreements between the protected area authorities and governments of Uganda, Rwanda and the DRC to support transboundary conservation (not to mention the dedication of national parks staff), have been important in maintaining a measure of success in conservation in the area through recent times of civil wars in the region (Plumptre et al., 2007b). However, the complexity of past human–environment interactions
and their role in shaping the current landscape is rarely acknowledged. The long-term effects of delimiting isolated forest fragments, with the aim of excluding all human activity within artificially created boundaries, are unknown. The framing of such approaches with a paradigm of retaining fragments of a ‘natural landscape’ does not take account of past human-induced modification of that landscape, as shown by evidence from this study.

A final issue raised by the new sediment-based evidence from the Virunga volcanoes relates to volcanic activity. The recent eruptive history of the volcanoes is not well-understood, although Mt. Muhavura is considered one of the most youthful (i.e. most recently formed) of the volcanic chain based on its relatively uneroded morphology. The continuous sedimentary record extending to ca. 3000 cal yrs BP indicates that this volcano has not erupted over at least the past three millennia (an eruption would have destroyed the sediments accumulating in the crater, as there are no side-cones to suggest an eruption from the flanks of Mt. Muhavura ever occurred). The cores extracted from the crater lake reached the base of sediment accumulation, indicating that the formation of the lake dates to around 3000 cal yrs BP. The length of time since the last eruption and the beginning of sediment accumulation in the crater lake on Mt. Muhavura is impossible to determine conclusively from evidence assembled to date, but the possibility certainly exists that the volcano was last active during the mid-Holocene. The sediment record from Mt. Gahinga extended beyond the range of coring equipment, indicating that a longer sedimentary archive than the 8000-year record extracted for this study exists at the site. This also confirms the older age of Mt. Gahinga compared to Mt. Muhavura. Evidence of tephra deposition at Mt. Gahinga may also provide information on former volcanism in the area.

The amount of tephra in the Mt. Gahinga record at ca. 5500 cal yrs BP is seven orders of magnitude greater than the amount at ca. 800 cal yrs BP. The equivalent tephra deposition event in the Mt. Muhavura lake at ca. 800 cal yrs BP was sufficient to affect aquatic conditions at the lake site. Given the generally low concentration of inorganic material throughout the highly organic Mt. Gahinga sediments, the tephra deposition at ca. 5500 cal yrs BP thus represents a relatively major volcanic event, which could possibly relate to an eruption of Mt. Muhavura.
Chapter 9 Conclusions

This thesis has sought to reconstruct the environmental history of a high-altitude area of the Albertine Rift, and in particular, to address three main research questions relating to the causes of environmental change during the Holocene. Multi-proxy sedimentary evidence from two crater sites in the Virunga volcanoes has been employed to answer the research questions, and the main results and findings of the research have been presented and discussed in the previous four chapters of this thesis. This chapter summarises the main findings associated with each of the three research questions, before reflecting on the limitations of this study and suggesting potential directions for future research.

9.1 Summary of main findings

A summary of the main findings related to the research questions is as follows:

Research question 1: To what extent have high-altitude sites in the Albertine Rift been affected by variations in monsoon activity?

This research has shown that climate changes linked to variations in monsoon activity during the mid- to late Holocene have been an important driver of environmental change, particularly vegetation change, both at high-altitude sites and surrounding areas of the Albertine Rift. Evidence from the pollen record from the Mt. Gahinga crater swamp has revealed several changes in the distribution and composition of vegetation during the mid-Holocene, and in particular in high-altitude (Ericaceous and Afroalpine) and montane forest plant communities, apparently driven by a shift towards increasing aridity. The timing of these changes in different vegetation types is asynchronous, and consists of a fire-related decline in Ericaceous vegetation between ca. 5200 and 4900 cal yrs BP, followed by a later increase in both high-altitude scrub and dry montane forest between ca. 4100 and 3700 cal yrs BP. This asynchronous timing of changes to high-altitude and montane forest vegetation types highlights the differential effects of climate change. An abrupt decline in fire-prone Ericaceous vegetation at high altitudes, concomitant with a macrocharcoal peak in the sediments, is indicative of a vegetation response to increased aridity, possibly in the form of greater seasonality (i.e. increased length of dry season).
Changes in the composition of lower montane forest vegetation on the Virunga volcanoes provide further support for a region-wide *Podocarpus*-rise, apparent at several montane sites in eastern Africa at ca. 4000 cal yrs BP. These changes show the widespread effects at all altitudes of monsoonally-driven changes in climate during the mid-Holocene, and provide new evidence for the differing sensitivities – in the form of asynchronous responses – of different vegetation types.

**Research question 2:** Do high-altitude sediment records from the Albertine Rift show evidence for increased human-induced environmental impact in recent millennia?

Evidence from both sites in the Virunga volcanoes shows a corresponding signal of increased human impact in the surrounding landscape at ca. 900 cal yrs BP. This comprises a significant (and sustained) forest clearance event, involving a reduction in lower montane forest taxa and increases in disturbance indicators. High-altitude (Ericaceous and Afroalpine) vegetation shows no major changes at ca. 900 cal yrs BP, thus effectively ruling out the possibility that widespread climate change was the main cause of vegetation change, and highlighting the role of human agency in causing forest clearance at this time – although a drier climate is likely to have affected other aspects of environmental change at this time. A further dimension to vegetation change during this period consists of a change in the composition and extent of *Hagenia*-dominated upper montane forest (which is currently widespread in the Virunga volcanoes), and has been linked to changes in the fire regime associated with increased human activity around the beginning of the second millennium AD. This highlights the extent to which human modification of the landscape within the last millennium has affected a wide range of vegetation types, even those previously thought pristine.

Results generated through the current research provide little evidence of major human-induced environmental changes prior to ca. 900 cal yrs BP, although human occupation of the area is known to have occurred long before this date. The lack of major human impact on the landscape until the most recent millennium is likely a reflection of low-intensity, possibly temporary inhabitation. The situation changed considerably following the onset of the Late Iron Age in the region, which marks the first occurrence of major human-induced
modification of the landscape. Furthermore, the increased role of human activity in influencing the landscape has continued throughout the second millennium AD, and has played a larger role in influencing environmental change than other factors, including non-anthropogenically induced climate change.

Research question 3: What have been the effects of variations in climate, fire, volcanic activity and human activity on high-altitude sites in the Albertine Rift?

This research has highlighted the complex interactions between different drivers of environmental change in the Albertine Rift. The research has also demonstrated that the relative roles of these drivers have changed during the Holocene. Up until ca. 900 cal yrs BP, evidence presented here indicates that the most important driver of environmental change was climate change linked to orbital variations, although climate-fire feedbacks were also important. The role of human activity, including human-induced effects on fire dynamics, has eclipsed natural forms of climate change as a major cause of widespread environmental change during the last millennium. Volcanic activity has not been found to have caused major changes in environmental conditions during the Holocene, but the deposition of tephra is likely to have influenced aquatic conditions and local vegetation in remote high-altitude ecosystems.

9.2 Limitations of the research

This research has aimed to reconstruct environmental change in the Albertine Rift based on sedimentary records from two high-altitude sites in the Virunga volcanoes. As with any sediment-based study, the extent to which a small number of sites reflect widespread regional change must be considered. The separation of localised and regional environmental change is most effectively addressed by utilising evidence from many sites over a large area. This problem has been minimised as much as possible within the constraints of this study by using two sites located at different altitudes within the Virunga volcanoes. Furthermore, the location of these two crater sites at an altitude above the current area of direct human modification of the landscape, and the similarity of the long-distance pollen component in the sediment records, provides an indication that these sites are reflecting environmental changes extending beyond their direct catchment areas.
One key assumption in the interpretation of the pollen diagrams from Mt. Muhavura and Mt. Gahinga is that the relative area from which long-distance pollen has been exported up to the crater sites has not changed appreciably during the period represented by the sedimentary records. A further assumption is that the mechanisms of pollen production and dispersal by the parent plant, and the taphonomic processes responsible for the incorporation of pollen into the sedimentary records, have also remained constant over time. There is no evidence to indicate that either of these assumptions do not hold: the ratio between long-distance and high-altitude pollen types remains similar throughout both pollen records, except for a slight change at ca. 900 cal yrs BP that can be attributed to the significant reduction in *Podocarpus* pollen exported to the sites. The wide uncertainty ranges associated with the age models for some parts of the sedimentary sequences, particularly between ca. 1000 and 500 cal yrs BP in the record from the Mt. Muhavura crater lake, mean that separating the timing of changes in different proxy measurements is somewhat problematic. The measurement of several proxy indicators of environmental change at different resolutions also leads to problems in comparing the timing of associated features. Increasing the resolution of some proxy measurements (notably pollen) with a specific focus on sediment depths associated with changes in charcoal and tephra values, would help to address this problem, although the precision of the age model estimates may still cause problems in identifying the exact timing of environmental changes.

### 9.3 Future directions

The findings of this thesis highlight several useful avenues for future research. The usefulness of high-altitude sites in the Albertine Rift as sentinels for detecting past environmental change in the region has been highlighted. Additional sediment-based research, both at other sites in the Virunga volcanoes and elsewhere in the Albertine Rift, would serve to provide further details on the extent and timing of the environmental changes identified in this study. A transect of sediment cores covering a range of altitudes and latitudes in the Albertine Rift, and a high-resolution focus on the periods of change between ca. 5500 and 3500 cal yrs BP, and between ca. 1000 and 700 cal yrs BP, would
provide additional details on the relative timing of major climate- and human-induced environmental change throughout different altitudes within the Albertine Rift.

Geochemical analysis of the tephra contained in the core sequences may enable the identification of the volcanic source area, and would also provide the potential to use the tephra horizons as correlative marker horizons in sedimentary records from elsewhere in the region. Investigations to establish a late Quaternary tephrochronological framework for eastern Africa are currently in their early planning stages (C. Lane, pers. comm.), and the tephra contained in the sediments from the crater sites on Mt. Muhavura and Mt. Gahinga is an indication of the excellent potential for such studies.

The sedimentary record at the Mt. Gahinga crater swamp is likely to extend back further than the ca. 8000-year record extracted for this study. Extraction of a deeper core sequence from this site may extend the record back to the late-glacial period, or even the last glacial maximum. This would provide exciting potential for the reconstruction of environmental change over a much longer time-scale, and would enable an assessment of the main driving factors of environmental change in the Albertine Rift throughout the late Quaternary.
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